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TROPICAL CYCLONE GENESIS in the WESTERN NORTH PACIFIC

by

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MAY 1975

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This paper presents climatological statistics on the location, frequency, and environmental conditions associated with seasonal tropical cyclone genesis of the western North Pacific. Data summaries are stratified into annual and monthly averages for the 28-year period of 1946-1973. The Annual Typhoon Reports of the U.S. Navy and Air Force from the Joint Typhoon Warning Center in Guam for the years 1949-1973 were used as the data source. It is observationally shown and physical reasons are given why seasonal cyclone genesis frequency is related to the product of the seasonally averaged parameters of: 1) low level ...

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20. Abstract (continued)

relative vorticity, 2) the Coriolis parameter, 3) inverse of the vertical shear of the horizontal wind from lower to upper troposphere, 4) an ocean thermal parameter related to the sea surface temperature and the sea surface minus 200 feet sea temperature, 5) moist stability from the surface to 500 mb, and 6) middle tropospheric relative humidity. A western North Pacific seasonal forecast potential of cyclone genesis frequency is derived. This forecast potential very well specifies on a seasonal basis the location and frequency of tropical cyclone genesis in this region.

TABLE OF CONTENTS

1. BACKGROUND AND PURPOSE

It is important that the best possible tropical cyclone climatology be made available to the operational personnel of the western North Pacific. This research has been organized with the purpose of imparting maximum climatological and statistical information to operational users so that more accurate planning and forecasting decisions can be made regarding tropical cyclone genesis. This paper attempts to establish a physical basis for understanding the processes of tropical cyclone genesis in the western North Pacific and of the variation in the seasonal climatology of genesis location and frequency. This is accomplished through an integration of our present observational information and through consideration of the likely relevant physical processes.

The first part of this paper presents statistical information on the seasonal frequency of tropical cyclone genesis and discusses what the author believes are the major physical requirements of genesis. The second part of this paper shows how these hypothesized genesis requirements are specified by the product of six seasonally averaged meteorological parameters. The last part of this paper shows how well the product of these seasonally averaged parameters is related to cyclone genesis location and frequency. A cyclone genesis forecast index is proposed.

2. GENESIS STATISTICS

Figure 1 gives the geographic location of the initial detection points of tropical disturbances which later became typhoons or tropical storms during the period of 1946-1973.[1] Figure 2 shows the geographical points where storms first reached typhoon intensity. Boxes surrounding 90 and 50 percent of these points and the centroid of all points is indicated by the 'C'.

Figure 3 shows the initial locations of the centers of satellite observed trade-wind mesoscale cloud clusters (3^o to 5^o latitude on a side) which later became typhoons during the seven years of 1967-1973. Note that the centroid of the initial cloud disturbance areas, as determined by the satellite (point C) is approximately 4^o of longitude further east than the centroid of the initial detection points of disturbances (point B) determined by the Joint Typhoon Warning Center (JTWC) from conventional information. As was expected the satellite data relocated farther east the initial detection points of disturbances from which tropical cyclones form. The centroid of the points of initial typhoon intensity (point A) indicates that it takes the initial cloud disturbance an average of approximately 11^o of longitude (or 2-3 days) movement to reach typhoon intensity.

Figures 4 through 11 portray the monthly (December through March and April-May are each treated as one period) locations of the initial

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1 Tropical cyclones are divided into two categories based on their maximum sustained low level winds: 1) tropical storms with speeds between 34 and 63 knots; and 2) typhoons with speeds greater than 63 knots.

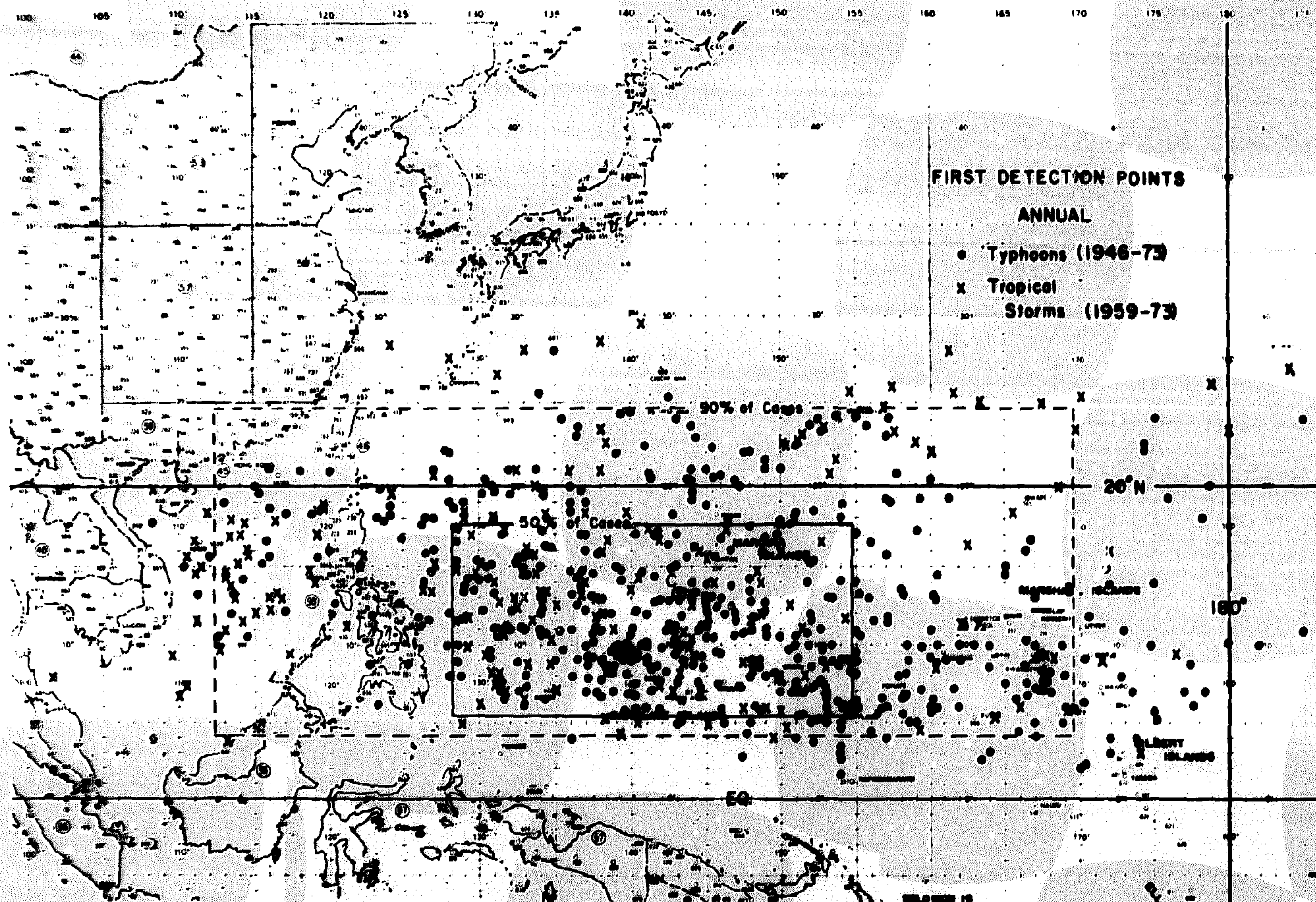

Fig. 1. Initial detection points of tropical systems which later grew to typhoon intensity (solid dots) for years 1946-1973 or of lesser intensity cyclones (x's) for the years 1959-1973. Solid and dotted lines outline 50 and 90 percent of all cases respectively. C denoted centroid of all points.

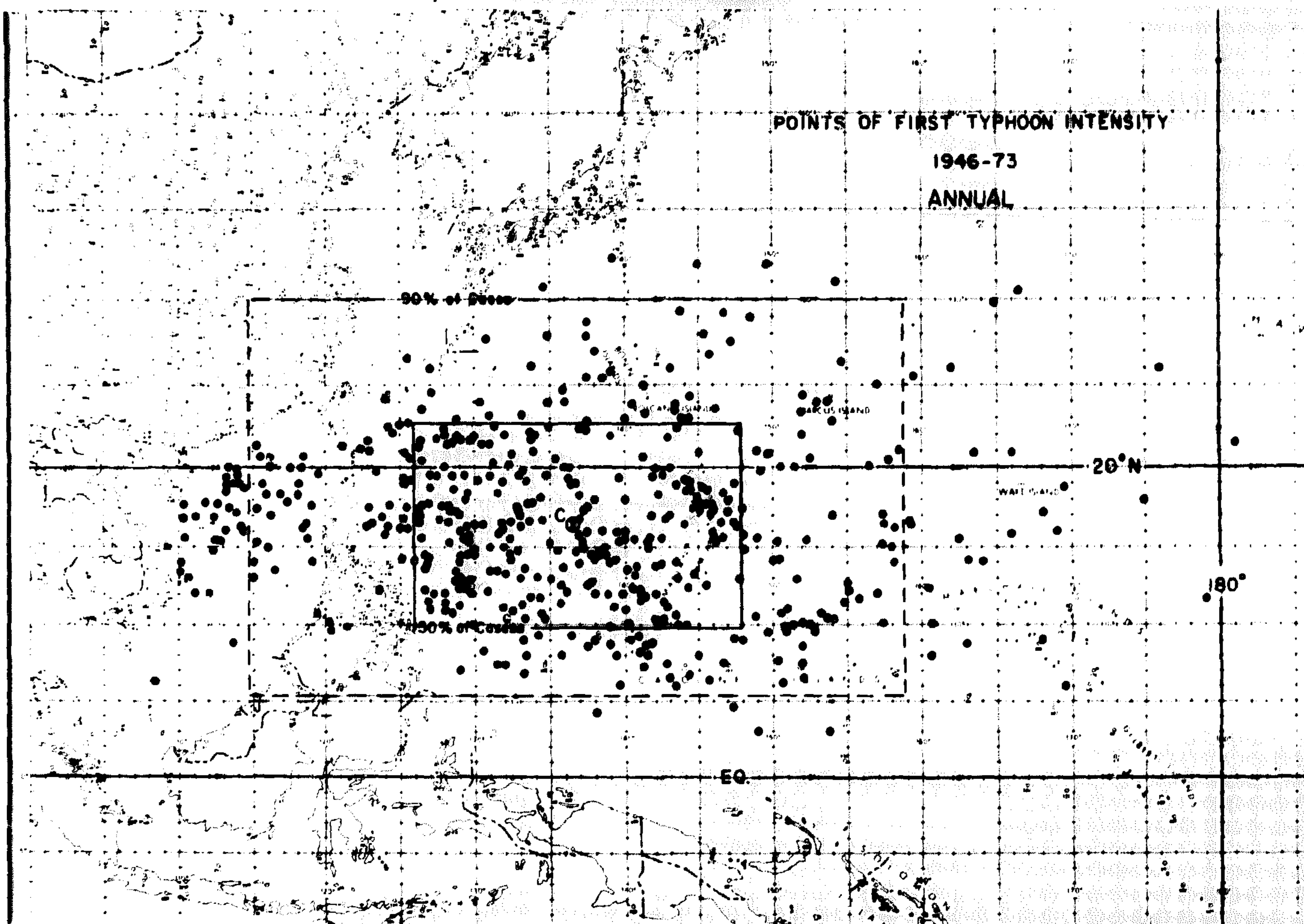

Fig. 2. Reconnaissance determined points of first typhoon intensity for the years 1946-1973. Solid and dotted lines outline 50 and 90 percent of all cases respectively. C denotes the centroid of all points.

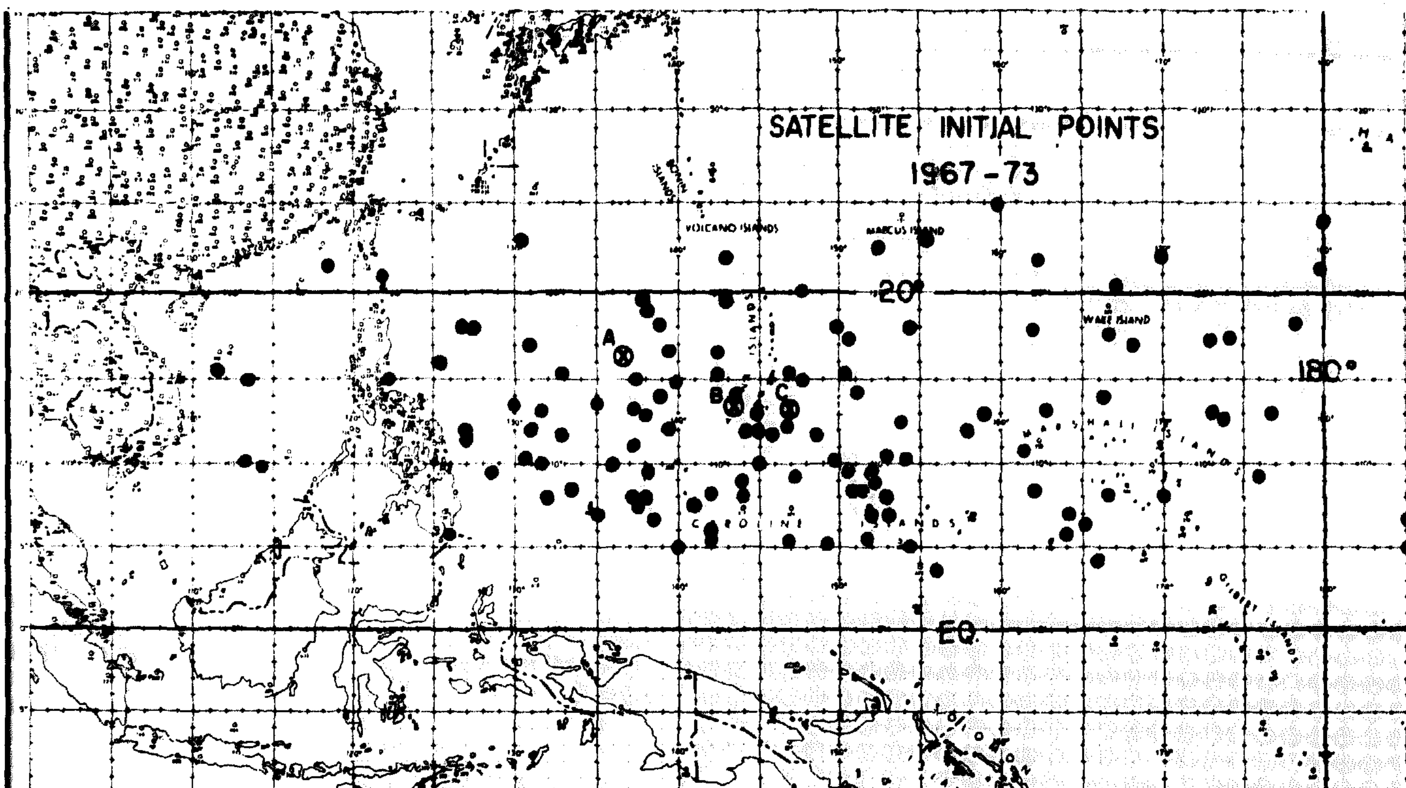

Fig. 3. Satellite determined initial typhoon and tropical storm origin points for the last seven year period when satellite data has been available. The A is the centroid of all the 1946-73 JTWC initial typhoon intensity points, B is the centroid of all the JTWC first detection points for 1946-73, and C is the centroid of all satellite determined initial detection points for 1967-73.

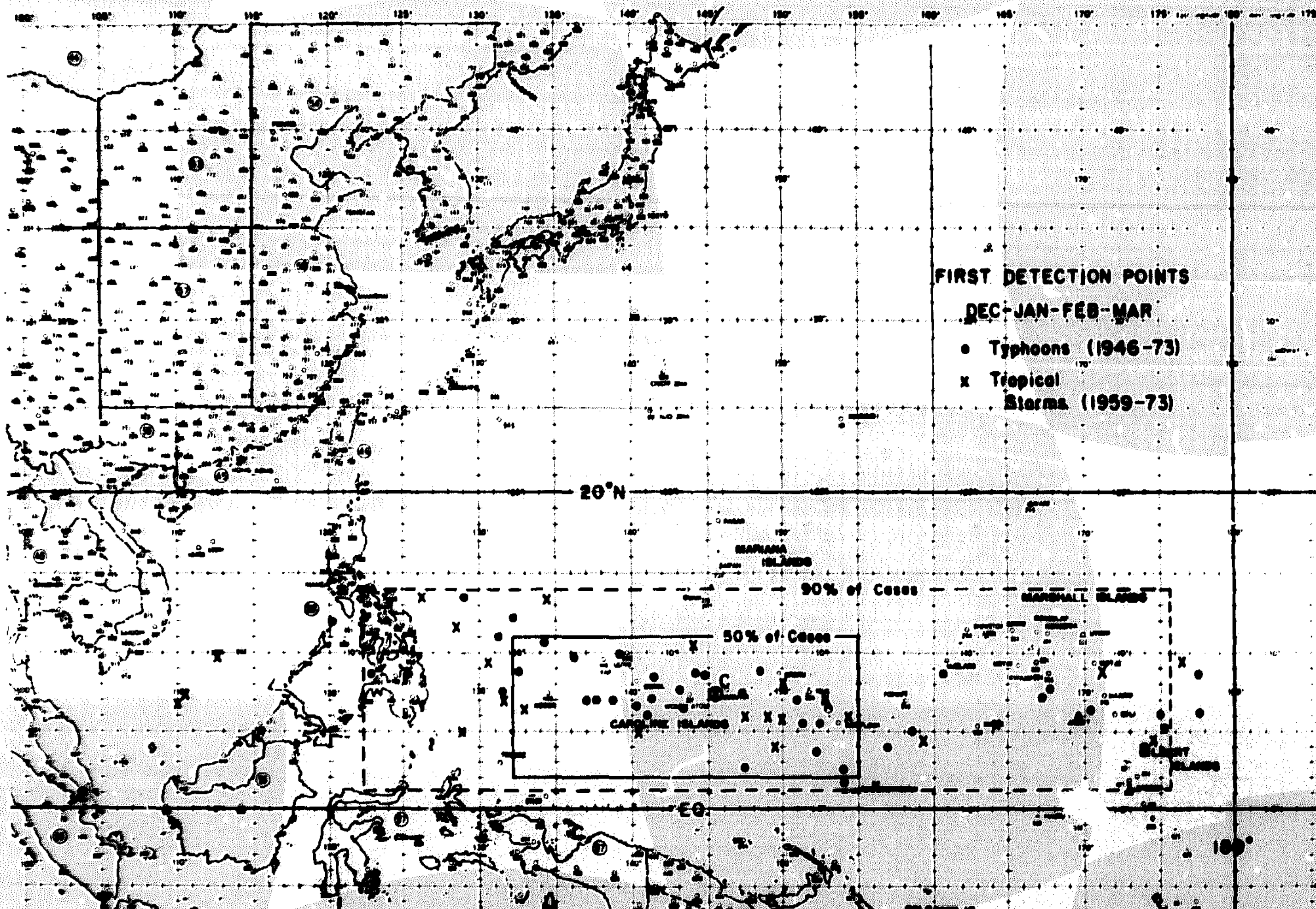

Fig. 4. December-January-February-March initial detection points of tropical systems which later grew to typhoon intensity (solid dots) or to cyclones of lesser intensity (x's). Solid and dotted lines outline 50 and 90 percent of all cases respectively. C denotes the centroid of all points.

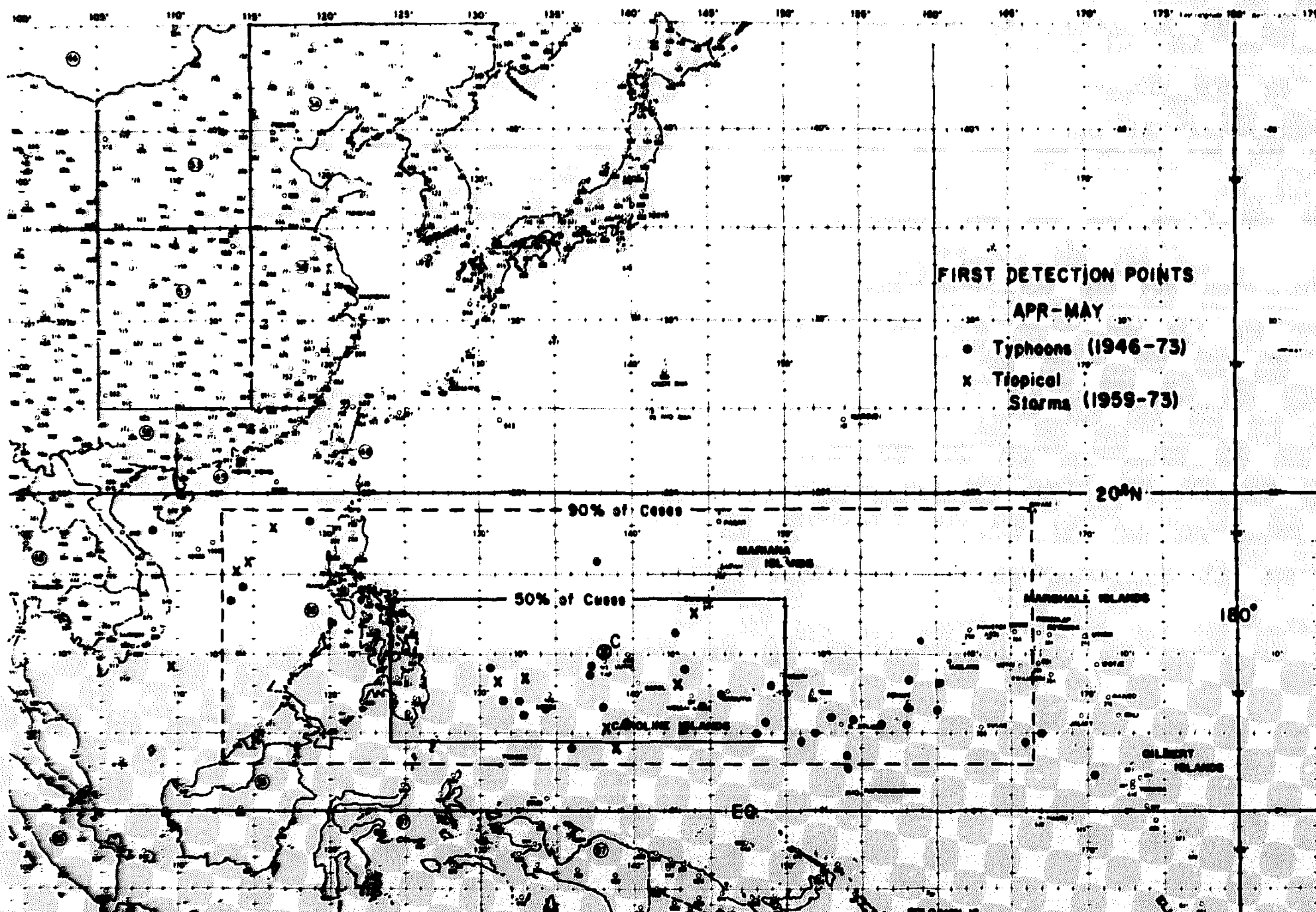

Fig. 5. Same as Fig. 4 but for the months of April-May.

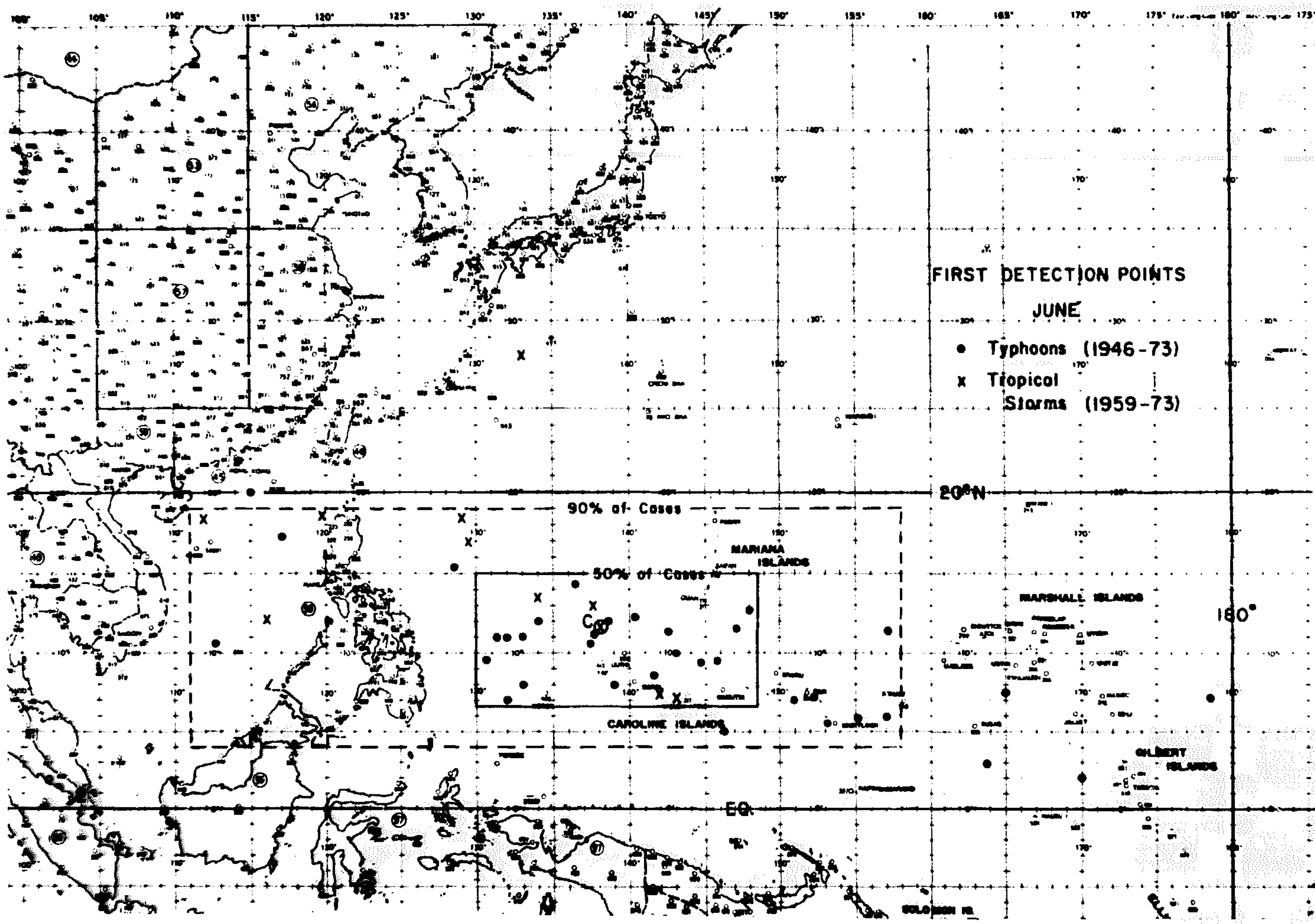

Fig. 6. Same as Fig. 4 but for the month of June.

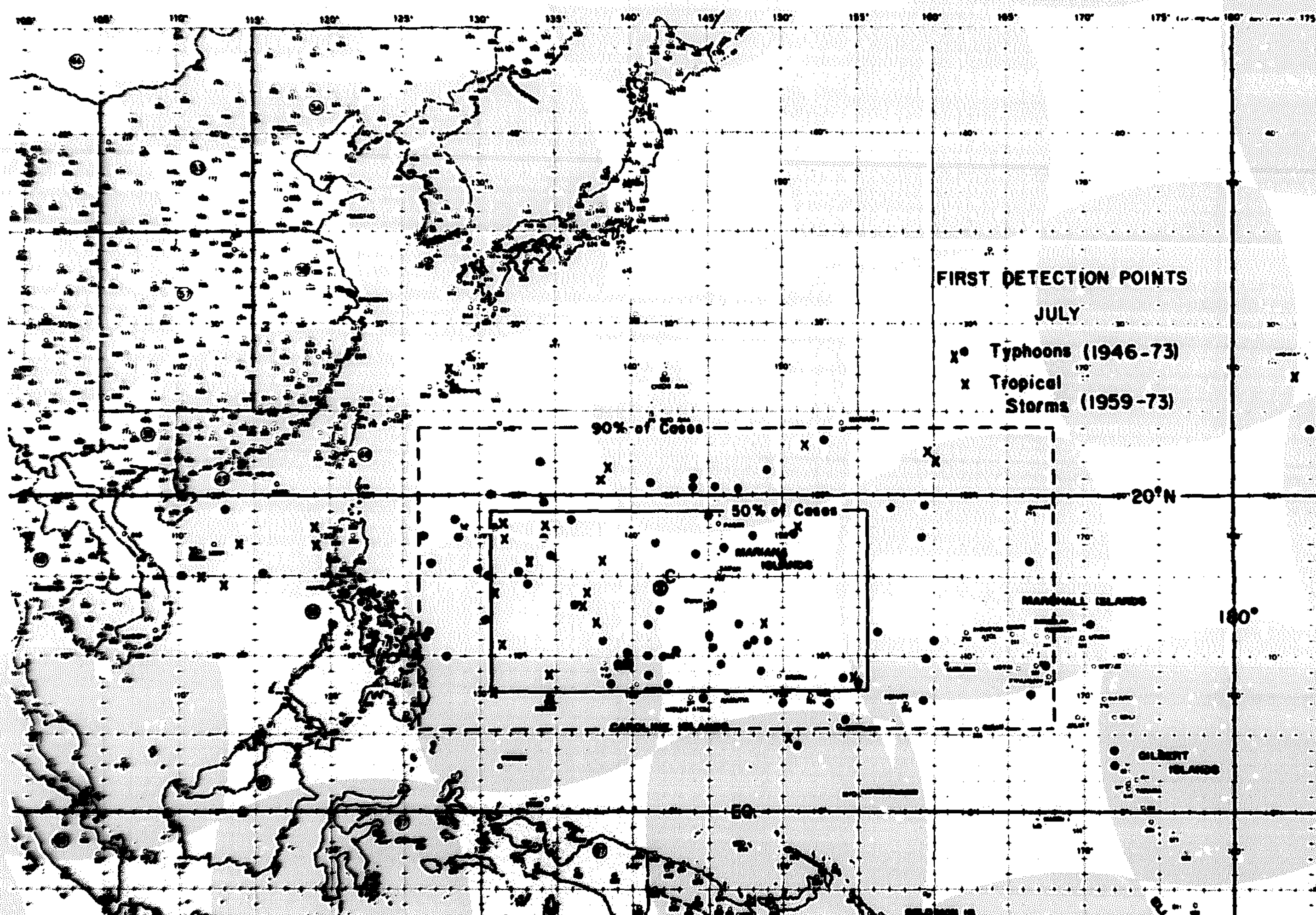

Fig. 7. Same as Fig. 4 but for the month of July.

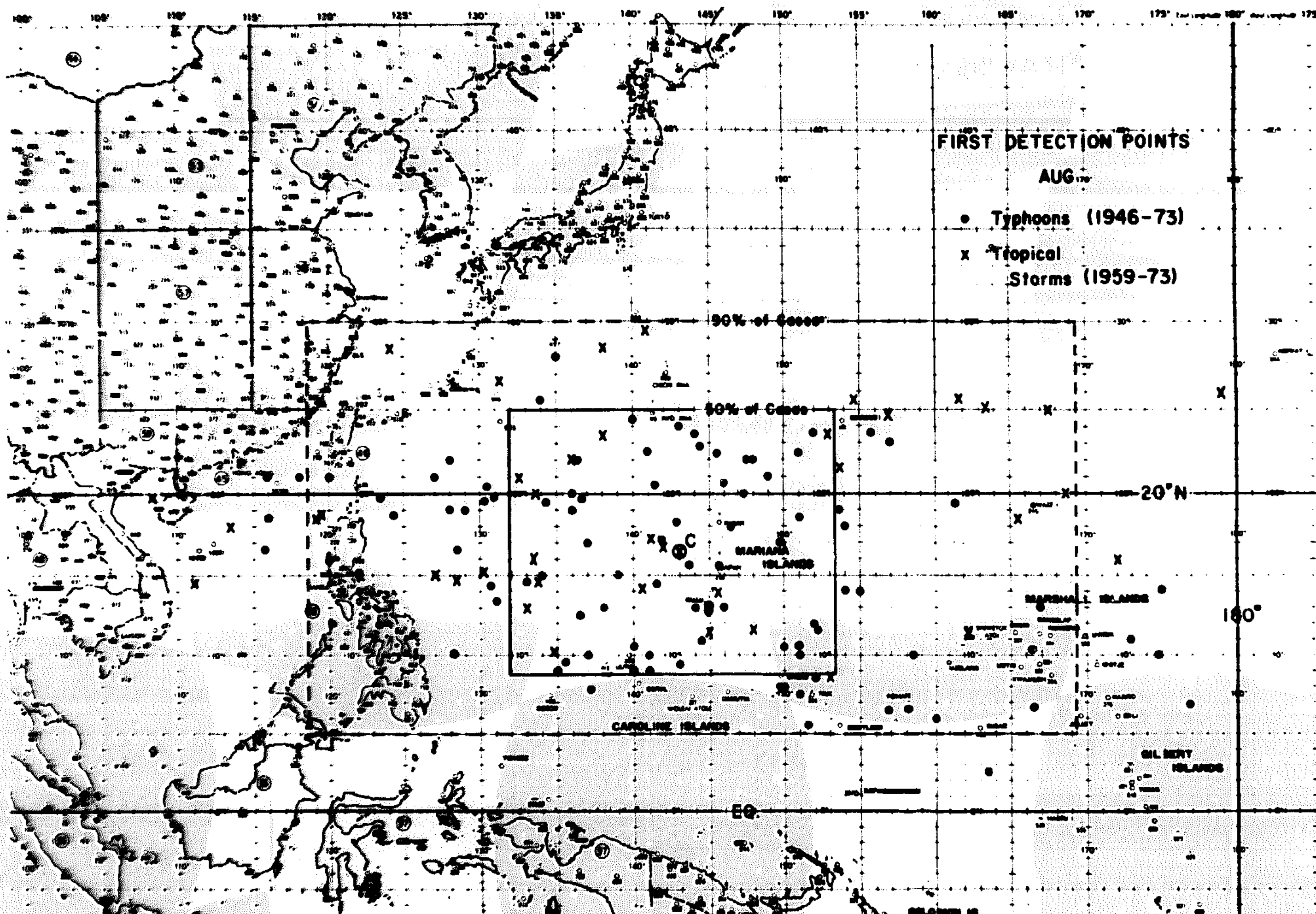

Fig. 8. Same as for Fig. 4 but for the month of August.

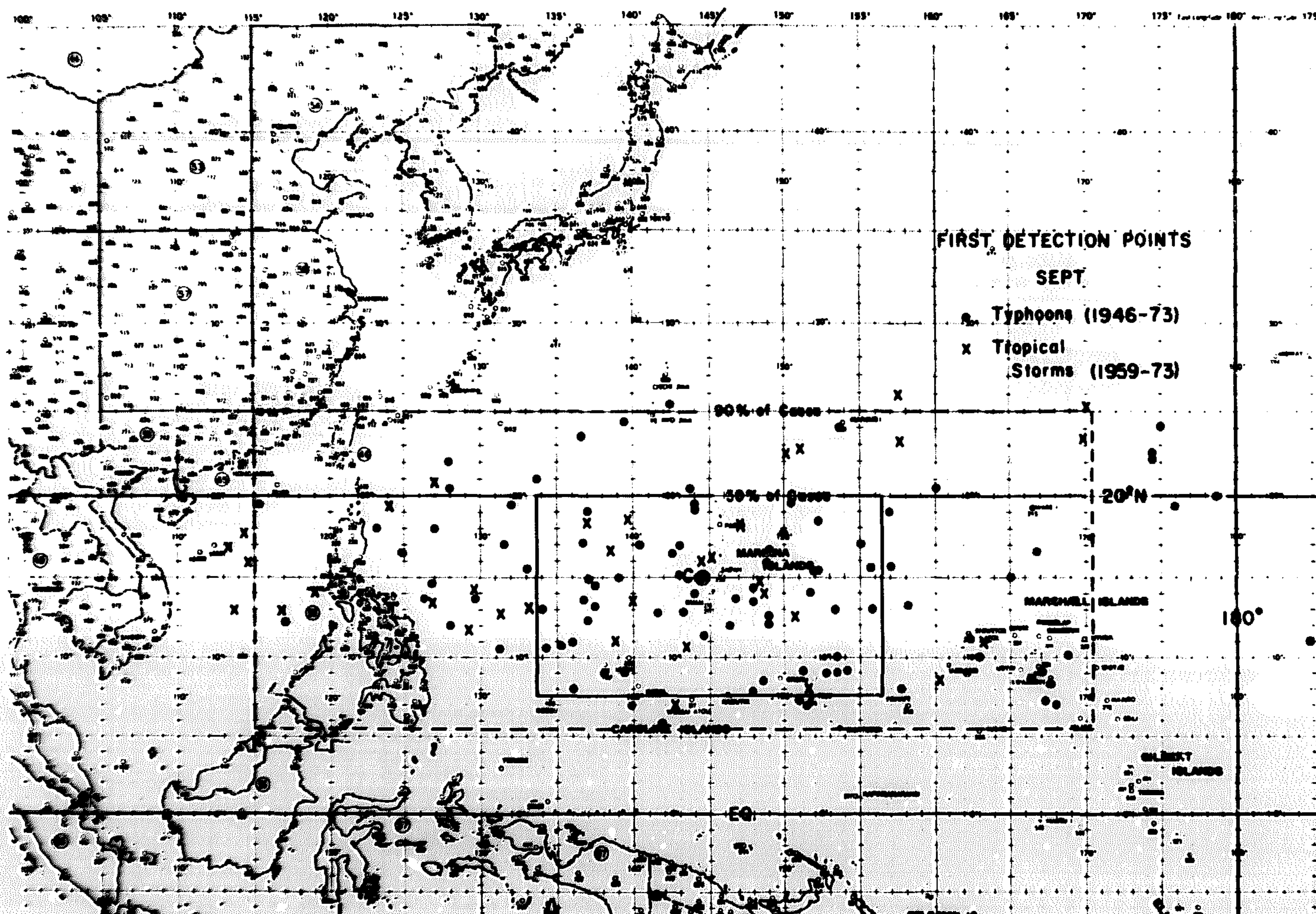

Fig. 9. Same as for Fig. 4. but for the month of September.

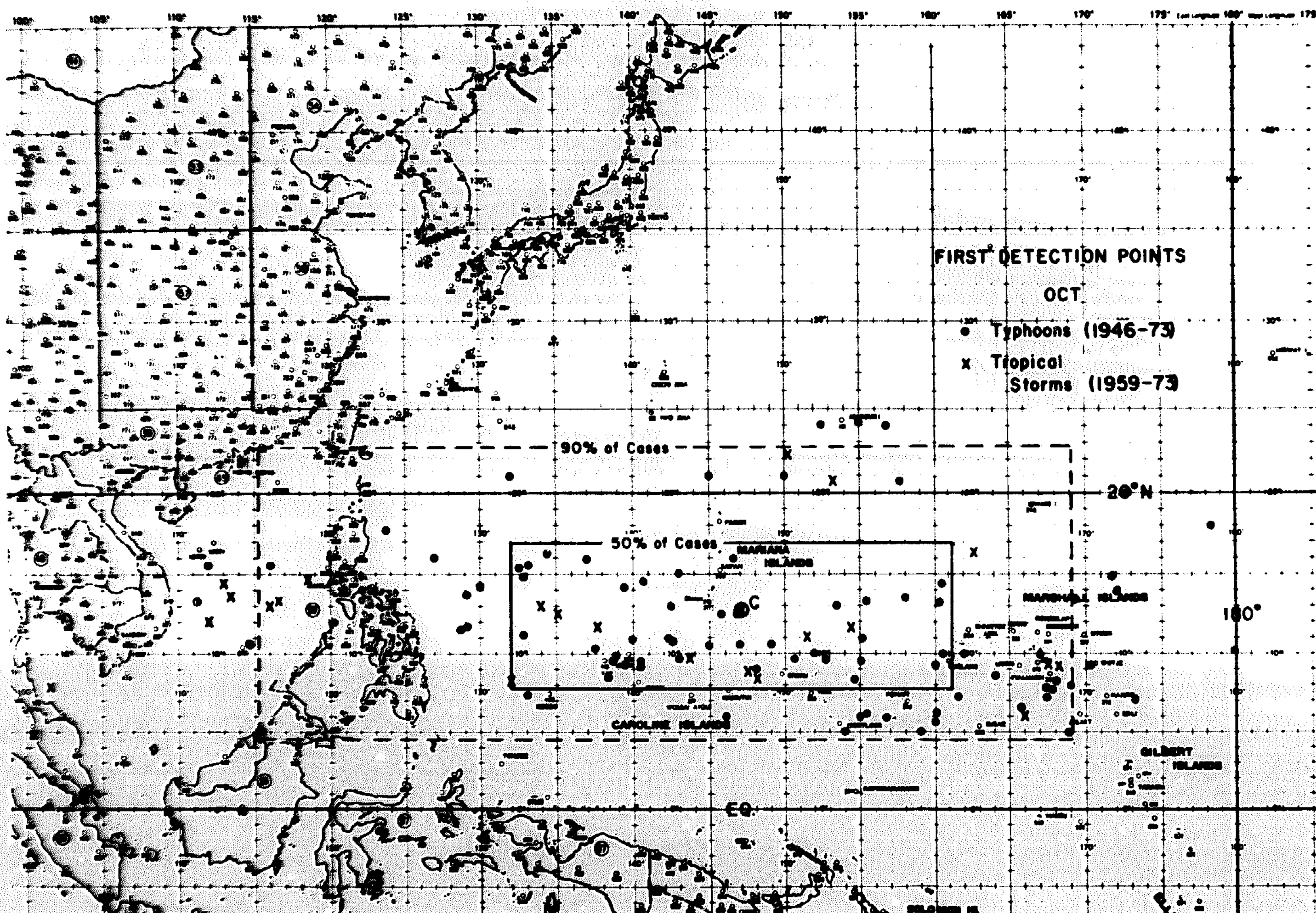

Fig. 10. Same as for Fig. 4 but for the month of October.

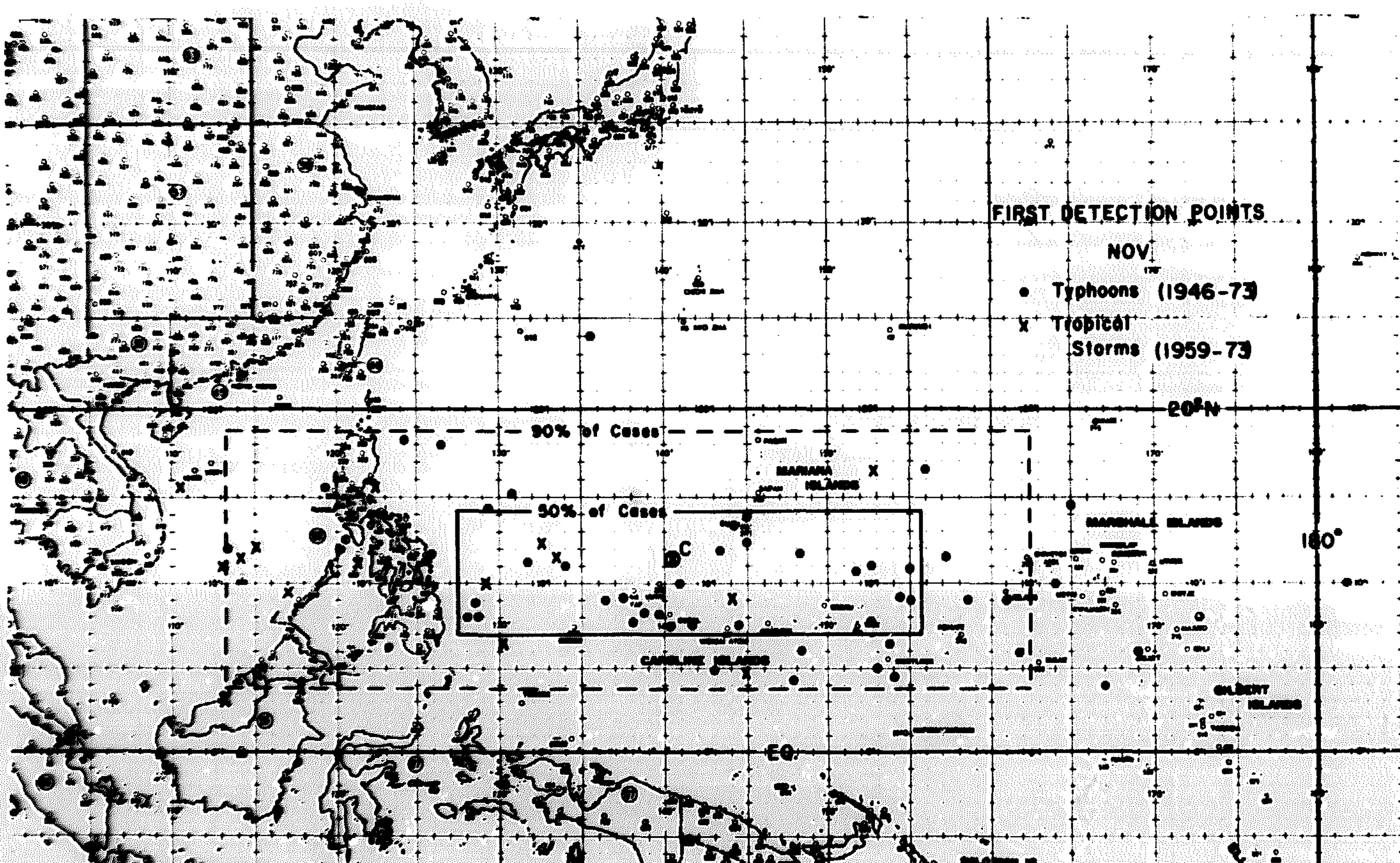

Fig. 11. Same as for Fig. 4 but for the month of November.

cloud systems or wind perturbations from which tropical cyclones later grew, as determined from 'TWC data. Solid dots represent tropical cyclones which reached typhoon intensity, and x's represent tropical cyclones which reached only tropical depression or storm intensity. Because of earlier period cyclone intensity determination difficulties, the lesser intensity or tropical storm data is felt to be representative only since 1959. Information on these latter intensity disturbances is presented only for this latter period. The 'C' denotes the centroid of all points. Little difference can be seen in the initial detection points of the two cyclone groups (x's and dots).

Because of observational limitations and the fact that the maximum wind speed of many tropical cyclones is near the 64 knot speed requirement, one cannot with certainty determine the exact number of typhoons and tropical storms for each season. If such classification errors are largely random (as we believe), then one can say that there were approximately 19 typhoons and 9 tropical storms per year in the western North Pacific in the last decade (1964-1973). Table 1 shows an updated yearly breakdown of typhoons vs. tropical storms since 1946 as given in the report of Hodge and McKay (1970). Note that the data of the last 15 years has shown an average typhoon and tropical storm increase per year of 3½ and 4 respectively. This is likely due to earlier observational deficiencies.

The strong correlation between cyclone genesis and upper tropospheric easterly winds is well established. A study of 10 years of U.S. Navy Fleet Weather Central, Joint Typhoon Warning Center (Guam), Annual Typhoon reports by the author has shown that the Guam analysis of 200 mb winds, as best they could be determined over incipient disturbances which later became typhoons, were from the NE, E, or SE in

Table 1

Number of Typhoons ($\geq$ 64 kt) and Tropical Storms (34-63 kt) Per Year for the Periods 1946-1958 and 1959-1973.

Year	Typhoons	Tropical Storms
1946	13	2
1947	19	8
1948	15	11
1949	14	8
1950	12	6
1951	16	1
1952	19	9
1953	17	6
1954	15	4
1955	19	3
1956	18	4
1957	18	3
1958	20	2
	215	67
Average (13 years)	16.5	5.2
1959	17	6
1960	19	8
1961	20	11
1962	24	6
1963	19	6
1964	26	13
1965	21	13
1966	20	10
1967	20	15
1968	20	7
1969	13	6
1970	12	12
1971	24	11
1972	22	9
1973	12	9
	289	142
Average (15 years)	19.2	9.5
OVERALL AVERAGE (28 years)	18.0	7.5

123 out of 158 cases. Cyclones seldom form in regions where the upper tropospheric wind is from the west.

Genesis is also possible at latitudes of 5-10° in the western North Pacific from January to March. At these latitudes winter trade wind strengths decrease sharply equatorwards and high values of low level cyclonic shear are present. Wintertime tropical cyclone genesis should not be unexpected at these low latitude locations.

With approximately 19 typhoons and another 9 tropical storms per year, the western North Pacific is the most active of the globe's eight tropical cyclone genesis regions (Gray, 1968). It accounts for approximately one-third of all the tropical cyclones that form. None of the other seven genesis regions experiences even half the number of tropical cyclones that occur in the western North Pacific. For instance, this region produces nearly three times as many tropical cyclones per year as does the Atlantic and Gulf of Mexico combined.

As previously discussed (op. cit.) about 80-85 percent of tropical cyclones originate in or just on the poleward side of the Inter-Tropical Convergence Zone (ITCZ). Most of the remainder ($\sim$ 15 percent) form in the trade winds at a considerable distance from the ITCZ but usually in association with an upper tropospheric trough to their northwest (Sadler, 1967a, 1967b, 1974). A few hybrid-type of warm-core cyclone systems ($\sim$ 1-2 percent) develop in sub-tropical latitudes along stagnant frontal zones. This latter type of cyclone is very atypical and not part of the discussion of this paper.

3. PHYSICAL REQUIREMENTS OF CYCLONE GENESIS

Over broad areas of the tropics where cyclones form, horizontal temperature gradients are practically non-existent. To understand tropical cyclone genesis one must understand how the tropical disturbance or tropical cloud cluster[2] is occasionally able to induce an upper troposphere enthalpy increase (or warming) and then concentrate this enthalpy increase over a small area of ∿ 200-300 km width. This upper tropospheric enthalpy increase produces thickness increase and (with the top of the system near 100 mb) small surface pressure decreases. Observations of early cyclone development show that this required enthalpy increase first takes place in the layers between 500-300 mb (Lopez, 1968; Yanai, 1961, 1963; Zipser, 1964) and is of the typical magnitude indicated in Fig. 12. This upper-tropospheric enthalpy increase is lacking in disturbances which do not develop into cyclones. It appears that the primary question of genesis rests with understanding how this upper tropospheric enthalpy increase (or warming) occurs and how it is concentrated and accumulated over the westward moving disturbance. This initial concentrated warming typically occurs on the side of the cloud cluster (LaSeur, 1962; Fett, 1964; Lopez, 1968), not at its center.

Cyclone Genesis Hypothesis. The author believes that the primary determinant for cloud cluster growth to cyclone strength rests with the characteristics of the upper tropospheric energy dispersion processes. Only rarely do the upper level wind conditions relative to a moving cloud cluster act to inhibit horizontal dispersion of energy and thereby

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2 In this paper the terms 'disturbance' and 'cloud cluster' (defined in GARP conference, 1968) are used synonymously. These clusters may be considered as being the 'pre-existing' disturbance from which Riehl (1954) states all tropical cyclones develop.

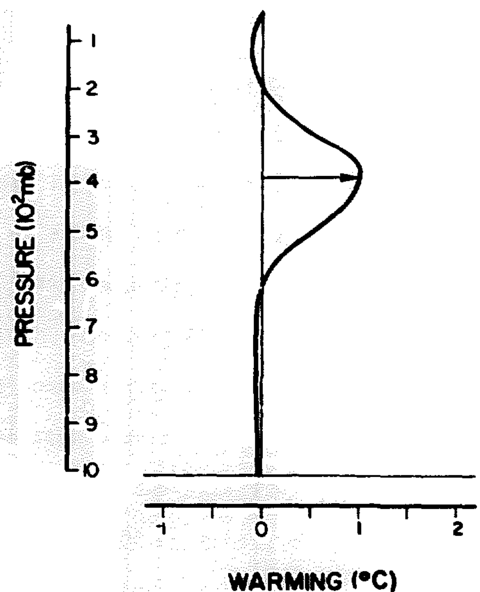


Fig. 12. Vertical distribution of the typical warming which is occurring in the usual 4° wide pre-cyclone tropical cloud cluster.

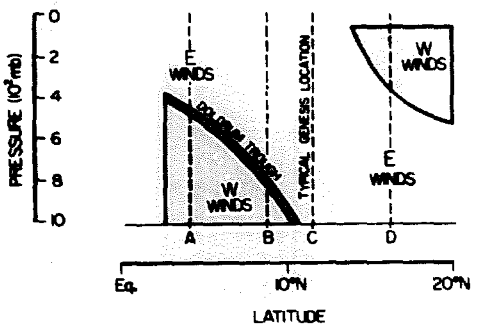
allow an upper level enthalpy accumulation within a moving system.

By horizontal energy dispersion the author means: 1) divergence, and 2) advection or ventilation. As the divergence profiles of the pre-cyclone and non-developing cloud clusters (Williams and Gray, 1973) are about the same, the horizontal energy losses from divergence cannot be a distinguishing feature of cluster growth. In addition, the magnitudes of the cluster divergences between 300-500 mb where the major warming occurs are not large, being near zero around 400 mb. Thus, the horizontal energy losses due to divergence are not felt to be the primary distinguishing factor to growth. Cloud cluster upper level advection or ventilation appears, on the other hand, to be fundamental in specifying cluster cyclone formation potential.

<u>Cluster Ventilation</u>. Gray (1968) has previously shown that ~85% of all tropical cyclones form within or just on the poleward side of a Doldrum Equatorial Trough or ITCZ. As shown in the north-south vertical cross-section of zonal wind in Fig. 13a this is an especially favorable place (location C) for small tropospheric vertical wind shear (also see third profile from the left of the lower diagram) and for 200-500 mb wind and cloud cluster velocity to be very similar. Designators A, B and D correspond to the typical locations of Monsoon, GATE[3] and trade wind cloud clusters relative to the Doldrum Equatorial Trough. These latter locations are places where the cluster 200-500 mb wind typically blows at a velocity considerably different than that of the cluster itself. Thus, the ventilation, or blow-through component of the zonal wind past these clusters is typically large. The lower diagram of this figure shows the typical vertical shear of the zonal wind (U) and the vertical distribution of the zonal wind component ($U - U_d$) relative to the typical moving (U_d) cloud cluster at these four locations. Most oceanic cloud clusters move at velocities of 6-8 m/sec (Chang, 1970; Wallace, 1970). All except the monsoon clusters move toward the west. The horizontal energy dispersion rates or ventilation rates between 200-500 mb for profiles A, B and D are much greater than for profile C.

In all but the pre-cyclone clusters the environmental air between 200-500 mb is typically blowing through the cluster in a time interval of but 10-20 hours. The cloud cluster induced warming (Fig. 12) which is occurring is advected or ventilated out of the cluster. An upper level

[3] GARP Atlantic Tropical Experiment conducted off West Africa between June and September in 1974.



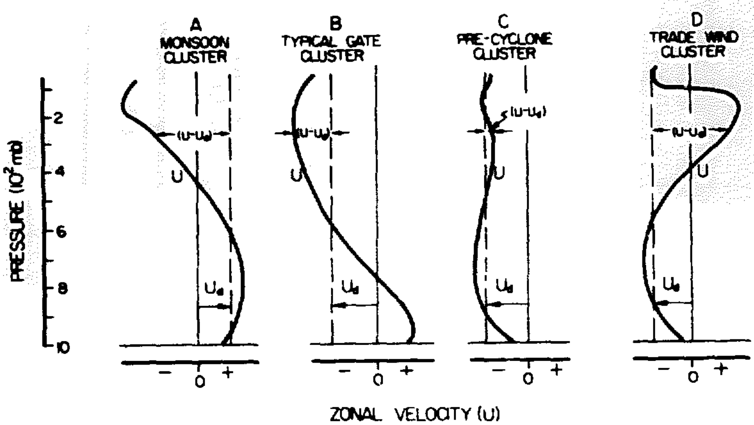


Fig. 13a. North-south cross-sections of the typical locations of various classes of tropical cloud clusters relative to the doldrum equatorial trough and the usual zonal winds present with these systems (top diagram). The bottom diagram portrays the vertical distribution of typical zonal wind velocity (U) occurring with different A to D types of cloud clusters whose general location is specified in the top diagram. The usual zonal cluster velocity is designated U_d and the difference in cloud cluster and environmental wind velocity at any level is given by $(U - U_d)$.

energy accumulation cannot occur. In the pre-cyclone cluster, on the other hand, the 200-500 mb motion of the air is very close to that of the cloud cluster motion and upper level energy accumulation can occur. In general, it is only cloud clusters with this type of vertical wind profile which produce tropical cyclones. Figure 13b shows the vertical profile of the average zonal wind in the western North Pacific pre-cyclone and non-developing cloud clusters. Note that the 200-500 mb zonal wind for the pre-cyclone clusters is very close to the westward velocity of the cloud cluster (data from Williams and Gray, 1973).

Once a closed circulation has been formed in the 200-500 mb levels, the direct flow of air or ventilation through the cluster is stopped. The enthalpy increase within the upper levels can be contained and concentrated. Cyclone intensification should and, in most cases, does follow.

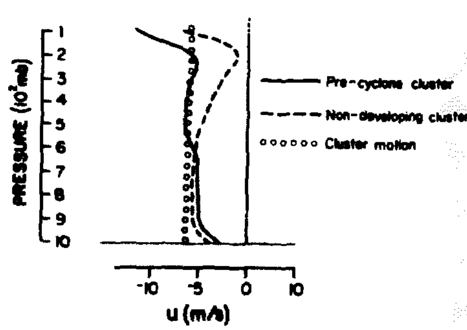


Fig. 13b. Vertical variation of the zonal wind velocity in the western North Pacific cloud clusters and the typical westward propagation of these cloud clusters. Note that the 200-500 mb zonal velocity of the pre-cyclone cluster is very close to that of the cluster westerly movement.

Summary. The formation of tropical cyclones whose circulation extends through most of the troposphere is a most difficult task for the tropical atmosphere to perform. It is seldom accomplished. The crucial problem in understanding cyclone genesis processes is that of understanding how the 200-500 mb levels can increase their enthalpy. It appears that this can be accomplished only if the upper tropospheric environmental winds and the disturbance from which the cyclone forms move with very much the same direction and velocity.

4. SIX PRIMARY GENESIS PARAMETERS

The author (Gray, 1968) has previously discussed the important role which large magnitude, low-level relative vorticity and small magnitude vertical shear of the horizontal wind play in determining regions of high tropical cyclone genesis frequency. Genesis also requires other favorable conditions to be present. A number of authors (Riehl, 1954, and much personal communication, 1957-1962; Fisher, 1958; Malkus and Riehl, 1960; Miller, 1964; Leipper, 1967; Perlroth, 1967, 1969; Leipper and Volgenau, 1972) have previously discussed the very important role which sea-surface temperature and high magnitude ocean thermal energy play in tropical cyclone existence. The necessary role of thermal buoyancy from the surface to middle levels for Cb convection has been discussed by Palmén (1948, 1957) and a number of other researchers. Other authors such as Dunn (1940, 1951) have observed the lack of tropical cyclone development near the equator and have hypothesized the importance of the earth's rotation in the genesis process.

In addition to these important genesis requirements the author has also observed the high frequency of cyclone genesis in regions where the seasonal values of middle level humidity are high (to be discussed).

The author believes that seasonal tropical cyclone frequency can be directly related to a combination of six physical parameters which will henceforth be referred to as primary genesis parameters. These parameters are:

1. Low-level relative vorticity (ζ_r)
2. Coriolis Parameter (f)
3. The inverse of the vertical shear (S_z) of the horizontal wind between the lower and upper troposphere, or $1/S_z$.

4. A Sea Temperature Factor (STF) of the sea temperature above 79°F between the surface and 200 feet ocean depth.

5. Vertical gradient of Θ_e between the surface and 500 mb ($\partial\Theta_e/\partial p$)

6. Middle troposphere relative humidity ($\overline{RH}$).

It will be shown that western Pacific seasonal cyclone genesis frequency is well related to the adjusted seasonal magnitude of the product of these six parameters, or

$$\begin{pmatrix}\text{Seasonal}\\ \text{Genesis}\\ \text{Frequency}\end{pmatrix} \propto (\zeta_r) \times (f) \times (1/S_z) \times (\text{STF}) \times \left(\frac{\partial\Theta_e}{\partial p}\right) \times (\overline{RH})$$

Previous authors have emphasized the importance for genesis of one or a few of these parameters. The author believes cyclone genesis is dependent on the magnitude of all of these parameters. The physical role which each of these parameters plays in the genesis process will now be discussed.

4.1 Role of Low Level Relative Vorticity

Tropical cyclones require a continuous low level import of mass, momentum, and water vapor. The magnitude of the import of these quantities in the Boundary (B) Layer (L), or surface to ∿ 1 km height, appears to be related in a significant way to the strength of the Ekman-type of frictional wind veering which is occurring. Over the tropical oceans it is observed (Gray, 1972) that this frictional veering of the wind averages between 10-15°. It has been further observed that the top of the layer of frictional wind veering does not increase near the equator as has been previously implied from Ekman theory, but rather is independent of latitude. The top of this layer appears to be largely determined by the degree to which mechanically and thermally driven air parcels in the mixed layer can penetrate into the stable layer just above. Over the tropical oceans the top of the veering level averages about 1-1.5 km while the top of the mixed layer is about 0.5 km.

Over the tropical oceans the average of the first km cross-isobaric or ageostrophic wind velocity, $\overline{V}_a$, is about 15 percent of the average B.L. total wind, or $\overline{V}_t$. Thus $\overline{V}_a \approx 1/7\overline{V}_t$, or in terms of the zonal $(\overline{u})$ and meridional $(\overline{v})$ wind components

$$\begin{aligned} \overline{u}_a &= -1/7\ \overline{v} \\ \overline{v}_a &= 1/7\ \overline{u}. \end{aligned} \tag{1}$$

Thus, the B.L. frictional convergence can be directly related to the B.L. average relative vorticity $(\overline{\zeta}_r)$ as

$$\begin{pmatrix} \text{B.L. Frictional} \\ \text{Convergence} \end{pmatrix} = -\left(\frac{\partial \overline{u}_a}{\partial x} + \frac{\partial \overline{v}_a}{\partial y}\right) = 1/7\left(\frac{\partial \overline{v}}{\partial x} - \frac{\partial \overline{u}}{\partial y}\right) = 1/7\ \overline{\zeta}_r \tag{2}$$

This type of frictionally driven convergence has been discussed by Charney and Eliassen (1949, 1964) and has been employed by many of the numerical modelers in simulating tropical cyclone growth. The association of surrounding low level vorticity with tropical cyclone frequency and cyclone dissipation over tropical water has also been discussed in project reports by Sartor (1968) and Wachtmann (1968).

It is well observed that tropical cyclones form only in regions of large positive low level ($\sim$ 950 mb) vorticity. The larger this low level vorticity, the greater appears the potential for cyclone genesis. Figure 14 gives seasonal values of 950 mb relative vorticity in values of 10^{-6}sec^{-1} plus 5 units (the data source is given in the appendix). The winter season is taken to be the months of January, February, and March; spring the months of April, May, and June, etc.

<u>Conclusion</u>: Other conditions being favorable and remaining constant, tropical cyclone genesis should be directly related to the magnitude of lower tropospheric relative vorticity.

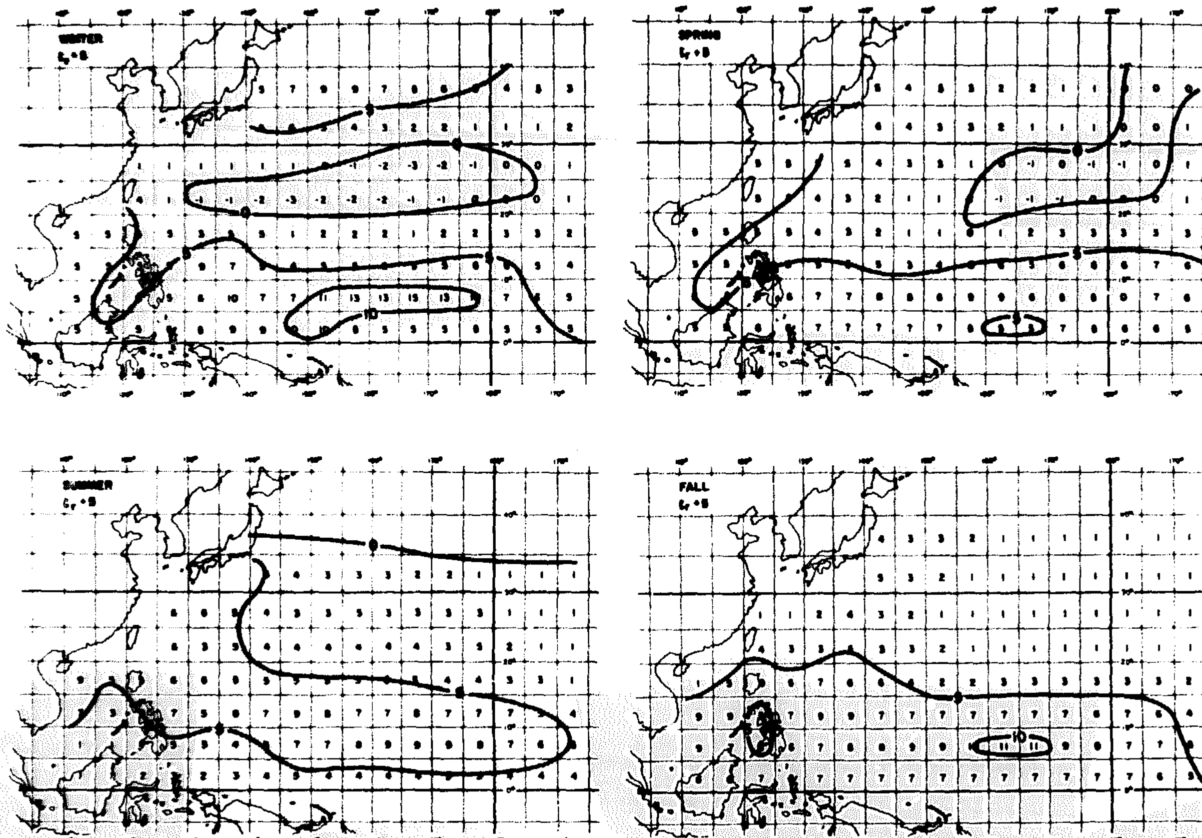


Fig. 14. Seasonal averages of 5° latitude-longitude square gradient level ($\sim$ 950 mb) relative vorticity (ζ_r) in units of $10^{-6}\mathrm{sec}^{-1}$ plus 5 units.

4.2 Role of Earth's Rotation (i.e. Coriolis Parameter)

Cyclones do not form within 4-5° of the equator. The influence of the earth's rotation would thus appear to be of primary importance. A number of previous researchers have stressed this relationship. Except for places where strong meridional flow and import of momentum from higher latitudes is occurring, winds on the equator are very weak. Geostrophic considerations dictate that pressure gradients near the equator be very weak.

Near the equator it is impossible to sustain a balance between cross-isobaric momentum acceleration and the planetary Boundary Layer (B.L.) frictional momentum loss when winds are higher than 3-5 m/sec. Assuming geostrophic conditions above the B.L., the ratio of the magnitude of B.L. kinetic energy generation ($f\,\overline{V}_g\,\overline{V}_a$) to frictional dissipation ($V_g \frac{\partial \tau_{xz}}{\rho \partial z}$) can be written as

$$\frac{f\,\overline{V}_g\overline{V}_a}{\overline{V}_g \frac{\partial \tau_{xz}}{\rho \partial z}} \tag{3}$$

where $\overline{V}_g$ is the geostrophic wind velocity in the B.L.

$\overline{V}_a$ is the average ageostrophic or cross-isobaric wind velocity through the B.L.

f is the Coriolis parameter

ρ is density, assumed constant through the B.L.

τ_{xz} is the usual stress representation in the B.L.

To keep the above ratio equal to one at low latitudes, only very small B.L. wind speeds are permitted. This is because dissipation is independent of f. B.L. momentum generation, on the other hand, is directly related to the pressure gradient which, except in highly curved flow, is a function of the Coriolis parameter.

Assuming the surface stress, τ_{xz_o}, is given by the usual formula $C_D \rho V_s^2$ (where C_D is a drag coefficient of the magnitude 1.3×10^{-3}, ρ is density, and V_s is the wind at 10 m height or ship deck level) and that this stress approaches zero at 1 km height (i.e., $\Delta Z = 1$ km - consistent with previous discussion), we can obtain an expression for the ratio of the momentum generation to the momentum dissipation, thus

$$\frac{f \overline{V}_g \overline{V}_a}{\overline{V}_g \cdot \frac{\partial \tau_{xz}}{\rho \partial z}} = \frac{f \overline{V}_a}{\frac{C_D V_s^2}{\Delta Z}}$$

Assuming that $\overline{V}_a$ through the boundary layer averages 1/7 V_s (see previous section) we obtain the final ratio

$$\frac{1/7 \; f \; \Delta Z}{C_D V_s} = K \frac{f}{V_s} \qquad (4)$$

where K is a constant.

For this expression to be constant V_s must be very small near the equator. Even when pressure gradients equivalent to curvature acceleration of 5 to 10 times that of geostrophic acceleration (fV_g) are allowed for, acceleration balance in the B.L. near the equator is possible only for conditions of very weak wind speeds.

This severe restriction on wind velocities near the equator should likewise restrict tropical cyclone formation. In the very early stages of cyclone formation the ratio of curvature to geostrophic acceleration is not overly large. It is assumed that cyclone genesis cannot proceed if B.L. velocities cannot be maintained.

Conclusion. Boundary layer dissipation is independent of latitude. Momentum generation, on the other hand, is latitude related. Cyclone intensification cannot occur if B.L. winds cannot be maintained. Other conditions being favorable and remaining constant, tropical cyclone genesis should be directly related to the strength of the Coriolis parameter.

4.3 Role of Tropospheric Vertical Wind Shear and Upper Level Ventilation.

As previously discussed by the author (Gray, 1968) the observational evidence clearly shows that tropical cyclones form under conditions of minimum vertical shear of the horizontal wind between the lower and upper troposphere. Because the primary cloud cluster enthalpy gain for cyclogenesis comes between the 200 and 500 mb levels, it is the horizontal winds at the upper levels which are most important in determining whether a cloud cluster will be able to accumulate enthalpy. The individual level cluster winds ($\mathbf{V}$) may be quite different than that of the cluster or disturbance velocity ($\mathbf{V}_d$). The magnitude of ($\mathbf{V} - \mathbf{V}_d$) determines the extent to which the cumulus induced upper level warming is advected, or ventilated, away from the disturbance. If this cloud cluster relative motion is small, an upper level enthalpy increase can occur. This will lead to gradual surface pressure decreases and cyclone formation. Once a cyclone has formed, the environmental motion relative to the system becomes very small and further enthalpy gain can be readily accomplished. If, however, the upper tropospheric flow relative to the motion of the disturbance is large, then 200-500 mb ventilation of enthalpy out of the cluster is too rapid to permit enthalpy concentration and accumulation. Pressure falls will not occur.

<u>Disturbance Ventilation and Enthalpy Accumulation</u>. Let $\left|\mathbf{V} - \mathbf{V}_d\right|_p$ represent the flow of the environmental wind through the cloud cluster at any level p. This is a measure of the non-divergent advection or ventilation of the disturbance. Enthalpy accumulation is a function of $\left|\mathbf{V} - \mathbf{V}_d\right|_p$ and the diameter (D) of the cluster. Cluster mean ventilation (Ven.) at any level p may be expressed as

$$(\text{Ven.})_p = \left|\mathbf{V} - \mathbf{V}_d\right|/2D \qquad (5)$$

and the Heating Retention Factor (H.R.F.) due to ventilation at any level p as

$$(\text{H.R.F.})_p = \left[1 - \left(\frac{\left|\mathbf{V} - \mathbf{V}_d\right|_p \Delta t}{2D}\right)\right] \qquad (6)$$

where Δt represents the interval of time over which the calculation is made. Stability of calculations requires that $\left|\mathbf{V} - \mathbf{V}_d\right|\Delta t / 2D$ be less than 1/2. This factor varies from 0 to 1.

Let H_p be the rate of sensible temperature accumulation $\frac{\Delta T}{\Delta t}$ at any pressure level p in the cluster due to warming processes or the cluster heating function. The net temperature accumulation with time from time t_o to t_x at any level is represented by $T_x - T_o$ and is given by

$$(T_x - T_o)_p = \sum_{t=t_o}^{t=t_x} (H)_p \; (H.R.F.)_p \; \Delta t \qquad \text{or} \qquad (7)$$

$$(T_x - T_o)_p = \sum_{t=t_o}^{t=t_x} \left(\frac{\Delta T}{\Delta t}\right)_p \left[1 - \left(\frac{|\mathbf{V} - \mathbf{V}_d|_p \; \Delta t}{2D}\right)\right] \Delta t$$

where

T_x = the final temperature at time t_x

T_o = the initial temperature at time t_o.

By assuming various values H, D, and $|\mathbf{V} - \mathbf{V}_d|$ one can calculate typical rates of enthalpy accumulation and see how these are related to D and $|\mathbf{V} - \mathbf{V}_d|$. Table 2 portrays how various values of $|\mathbf{V} - \mathbf{V}_d|$ and D influence the accumulation of enthalpy at an individual level over a period of three days if the heating function $(H)_p$ is assumed to be 1°C/day over the whole period. Note the large influence which ventilation has on reducing the disturbance enthalpy accumulation. If the heating occurs over a large area, the inhibiting influence of $|\mathbf{V} - \mathbf{V}_d|$ is much reduced. Thus, a cloud cluster with a very narrow heating area (or small value of D) is much less able to accumulate enthalpy.

If the assumed vertical enthalpy increase or heating rate of Fig. 12 is applied to clusters with various intensities of ventilation, the vertical distributions of temperature accumulation for a heating function of 250 km width are as portrayed in Fig. 15. The left diagram shows the temperature accumulation which would result if ventilation at all levels were zero. The center diagram shows the temperature accumulation which results with only weak ventilation and the diagram on the right the enthalpy accumulation which results with the usual cloud cluster. In the first case surface pressure falls of 4-5 mb would

Table 2

Accumulated Temperature Increase (^{o}C) vs. Time for Different Values of Disturbance $|V - V_d|$ and D and for a Heating Rate of $1^{o}C$/day. Warming Values Were Calculated From Eq. 7.

For Disturbance Warming of 100 km Diameter

$\|V - V_d\|$ in m/sec	t_o	12 Hours	1 Day	2 Days	3 Days
	Time (t)				
0	0	0.5	1.0	2.0	3.0
5	0	0.3	0.4	0.4	0.4
10	0	0.1	0.2	0.2	0.2
15	0	0.1	0.1	0.1	0.1

For Disturbance Warming of 250 km Diameter

$\|V - V_d\|$ in m/sec	t_o	12 Hours	1 Day	2 Days	3 Days
	Time (t)				
0	0	0.5	1.0	2.0	3.0
5	0	0.3	0.6	0.8	0.8
10	0	0.2	0.4	0.5	0.5
15	0	0.2	0.3	0.4	0.4
20	0	0.1	0.2	0.2	0.2

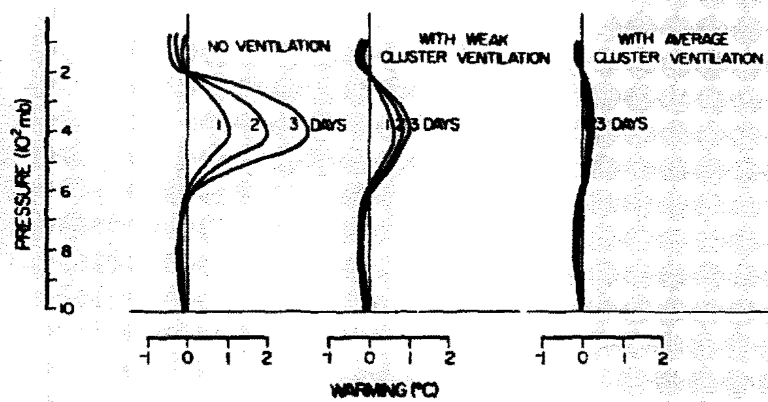


Fig. 15. Rate of temperature accumulation for 250 km wide cloud clusters with identical heating function of Fig. 12. but with different magnitudes of ventilation.

occur in three days. In the second case pressure falls of 1-2 mb occur after three days. In the third case the surface pressure decrease after three days is only ∿ 0.1 mb. In the first case a cyclone rapidly forms. In the second case a cyclone slowly forms. This is typical of the usual genesis situation. No cyclone formation is possible in the third case. Lopez (1968) has previously analyzed two cloud cluster cases of cyclone growth and non-growth and indicated differences in heating rates approximately equivalent to the two cases on the right of this figure.

The author believes that the major differences in realized net warming between the developing and non-developing cloud cluster is due primarily to upper level ventilation differences and not to heating function differences. Thus, the pre-cyclone cloud cluster and the usual cloud cluster have very similar rainfall rates, but, due to ventilation differences, their intensification potentials can be very different. Riehl (personal communications, 1957-1962) has often remarked on the lack of direct association of disturbance rainfall and disturbance intensification potential.

These ventilation differences, can, in general, be well represented by the vertical shear of horizontal wind between the low and upper troposphere. The tropical forecast offices in Miami (Simpson, 1971) and Guam (Atkinson, 1973) are presently using tropospheric vertical wind shear as a primary factor in forecasting cyclone genesis. Cyclones do not form when the 950 to 200 mb vertical wind shears are greater than about 10 m/sec or when the 200-500 mb wind velocities relative to the cloud cluster motion are larger than about 5 m/sec.

Conclusion. Other conditions being favorable and remaining constant cyclone genesis potential should be inversely related to the magnitude of the cloud cluster ventilation between 500 and 200 mb. This ventilation can be rather accurately expressed in terms of the vertical shear of the horizontal wind between 950 and 200 mb. Figure 16 gives western Pacific seasonal values of the magnitude of 950mb to 200 mb vertical shear of the horizontal wind in m/sec plus 5 arbitrary units.

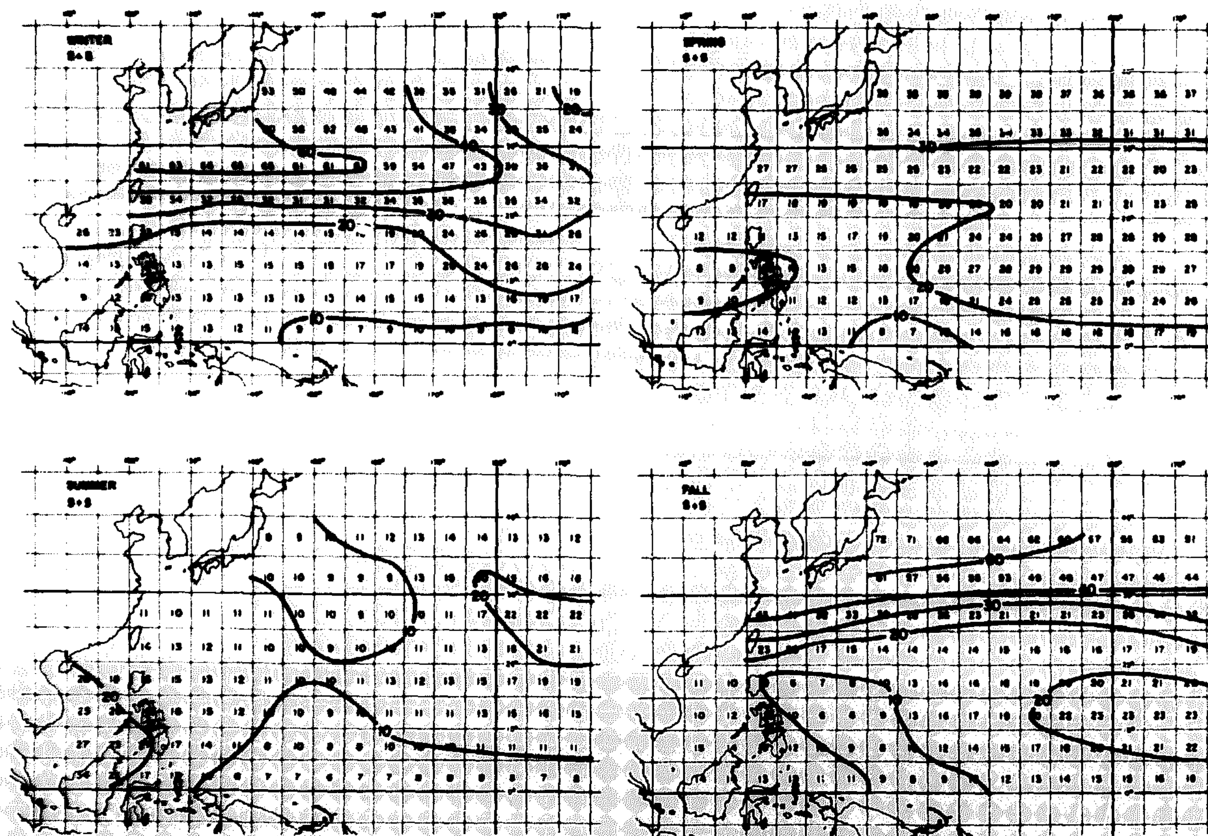


Fig. 16. Seasonal averages of 5° longitude-latitude square vertical shear of horizontal wind between 950 and 200 mb in units of m/sec plus 5 arbitrary units.

4.4 Role of Ocean Temperature

As previously discussed by Jordan (1964), Landis and Leipper (1968), Leipper (1967),Leipper and Jensen (1971), Leipper and Volgenau (1972), Heffernan (1972), and Perlroth (1967, 1969) tropical cyclones can have a profound influence on the temperature of the ocean over which they travel. The altered ocean temperature, in turn, can feed back and alter the character of the cyclone.

Leipper and his research group have indicated that the inner region of the average hurricane consumes about 4000 cal/cm^2 per day of ocean sensible and latent heat energy. In their study of the hurricane boundary layer Malkus and Riehl (1960) put this value at $\sim$ 3100 cal/cm^2 per day over the inner 90 km radius of the storm. Whitaker's (1967) calculations for Hurricane Betsy indicate a value of about 3700 cal/cm^2 per day. This large surface energy requirement of the hurricane precludes its existence over land or over the ocean if these transfers should somehow be suppressed. Brand (1971) has discussed how typhoons can weaken when they cross the track of a previous typhoon which has produced a cooler sea surface temperature due to upwelling.

The tropical cyclone not only gains energy from the ocean but its cyclonic wind circulation causes oceanic upwelling near the storm's center. The wind circulation produces a cyclonic ocean circulation which is not balanced by an equivalent upward and outward slope of water surface. The reduced air pressure at the storm's center may even cause a slight rising of the inner core ocean surface. Oceanic divergence is developed as a consequence. Upwelling occurs within and around the cyclone center (Black and Mallinger, 1971) as a result of this oceanic divergence. If the ocean has only shallow warm water, then a cold water upwelling will occur and the ocean will not be able to transfer its required energy to the air. Surface air temperature will decrease. The potential moist instability of the inner cyclone wall will also be much reduced. O'Brien and Reid (1967) and O'Brien (1967) have provided numerical confirmation of these central hurricane upwelling physical arguments. Leipper and his research colleagues have provided observational arguments.

Figure 17 shows the influence of reduced surface air temperature on surface values of Θ_e for a constant relative humidity of 85 percent. A very substantial decrease in the Θ_e gradient between the surface and 600 mb is noted for surface temperature decreases of only 2-3°C. Surface temperatures less than 26°C do not permit sufficient thermal buoyancy to sustain Cb convection. Palmén (1948, 1957) has earlier demonstrated a well defined cut-off of tropical cyclone activity with the 26½$^{\circ}$C ocean isotherm. Middle tropospheric values of Θ_e undergo much less daily, seasonal, and geographic variation than do surface values. Thus, potential moist buoyancy ($\frac{\partial \Theta_e}{\partial p}$) between the surface and middle layers is primarily influenced by the change of sea surface temperature.

Sea Temperature Factor. Leipper and Perlroth have indicated that the influence of hurricanes on the ocean underneath them can extend down to about 200 feet. Given the well established criteria of surface air temperature being above 26°C, or 79°F (Palmen, 1948) a Sea Temperature Factor (STF) is proposed and defined as

$$STF = \frac{T_{SST} - 79^{\circ}F}{1 + (T_{SST} - T_{200\ ft})} \qquad (8)$$

where T_{SST} = Sea Surface Temperature in $^{\circ}$F

$T_{200\ ft}$ = Sea Temperature at 200 feet depth in $^{\circ}$F

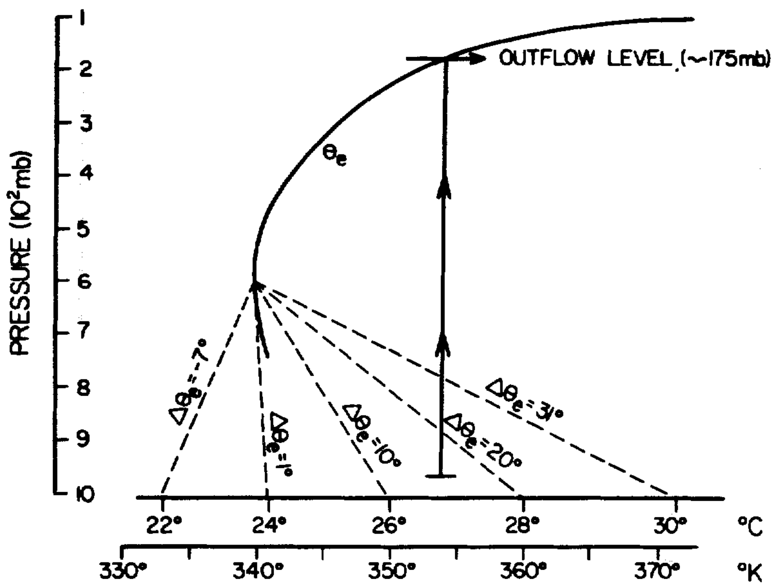


Fig. 17 Portrayal of how oceanic surface temperature changes can influence the Θ_e gradient between the surface and 600 mb under assumed constant surface humidity of 85 percent. Solid arrows portray the path of undiluted moist ascent from a boundary layer Θ_e value of 354°K to the level of maximum cyclone outflow.

This parameter combines the joint requirements for ocean temperature being above 79° F and for warm water to be present down to depths of 200 feet. Figure 18 portrays seasonal values of this factor.

Conclusion. Other conditions being favorable and remaining constant, tropical cyclone genesis should be directly related to the magnitude of the Sea Temperature Factor (as here defined).

4.5 Role of Surface to Middle Troposphere Θ_e Gradient

Except when there is deep cumulus convection to produce vertical coupling, the flow features of the tropical upper and lower troposphere appear to operate independently (Riehl, 1954). Tropical cyclone circulations, however, extend through the entire troposphere. Cyclones do not occur unless there is a well established vertical coupling of the lower and upper tropospheric flow patterns. Cumulonimbus convection acts as the primary mechanism for this vertical coupling.

Cb convection requires substantial ($\sim 10^{\circ}$K or more) decrease of Θ_e between the boundary layer and the middle troposphere (lowest values of Θ_e are typically at 600 mb, see Fig. 17). A number of other researchers such as Aspliden (1971) have discussed the characteristics of the vertical gradient of Θ_e over the tropical oceans with different types of convection. For convenience and following the author's previous paper (Gray, 1968) Θ_e gradients are taken between the surface and 500 mb. These values typically range between 15-20°K. Figure 19 gives seasonal values of this parameter plus 5 arbitrary units. Except in subtropical latitudes in winter, the values of this parameter are usually adequate for cyclone development. Daily parameter deviations are small and typically not well related to cyclone genesis potential.

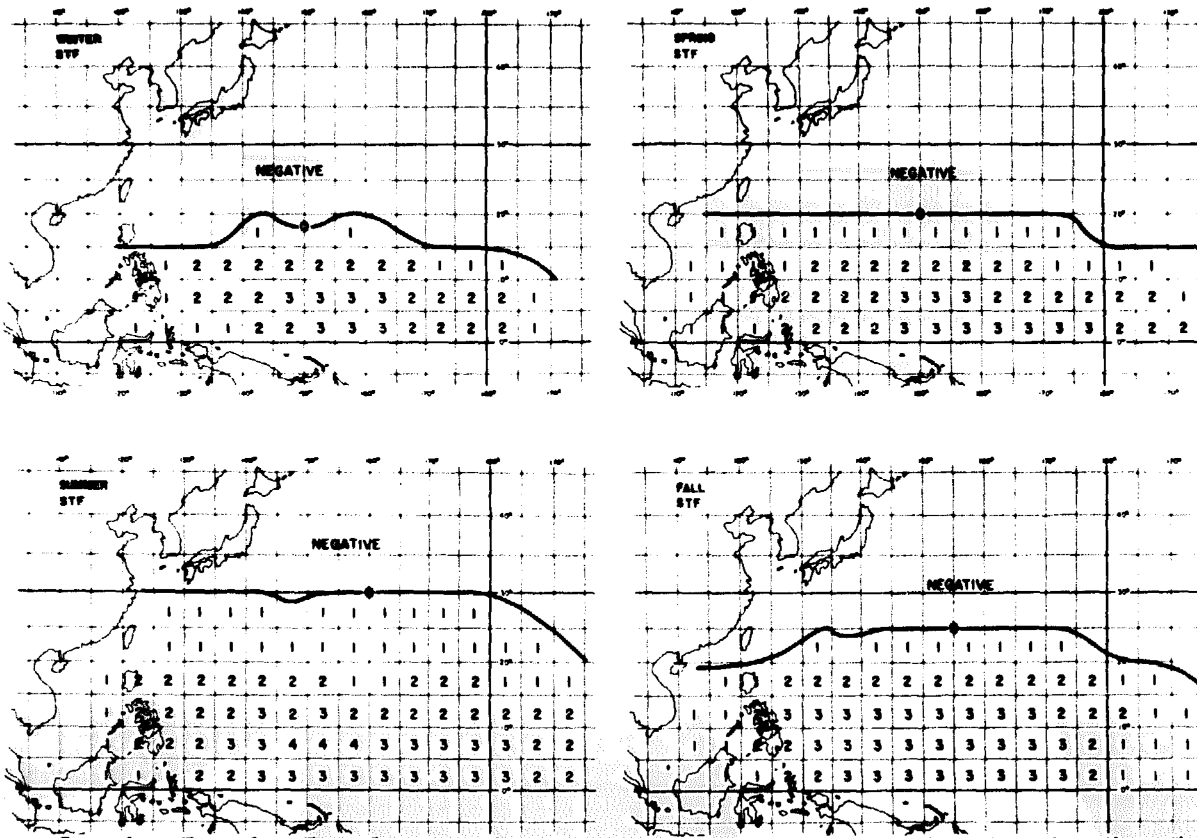


Fig. 18. Seasonal average of 5° square Sea Temperature Factor (STF) or $(T_{SST} - 79°F)/1 + (T_{SST} - T_{200\ ft})$ where T_{SST} and T_{200} are the sea temperature at the surface and at 200 ft respectively, in °F.

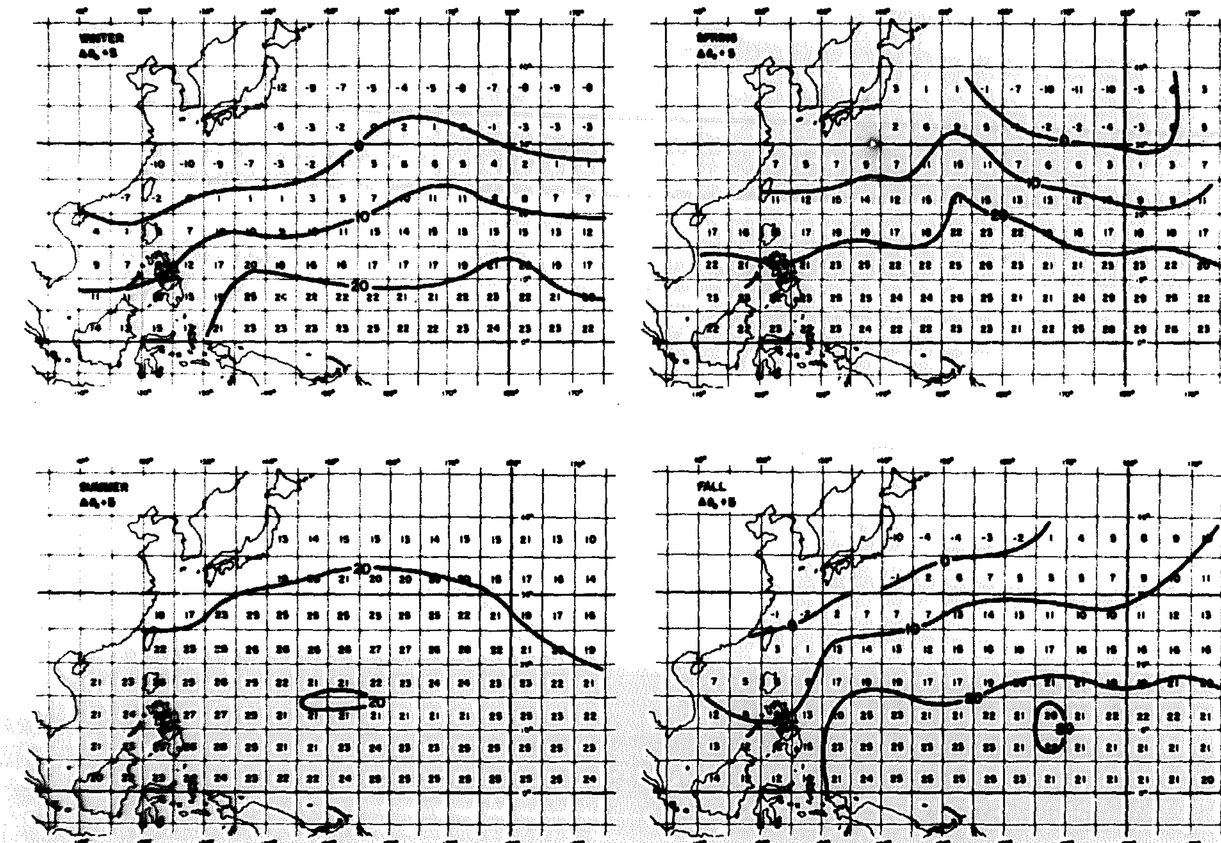


Fig. 19. 5^{o} square seasonal values of the vertical gradient of Θ_e ($\Delta\Theta_e$) between the surface and 500 mb in units of ^{o}K plus 5.

Conclusion. Other conditions being favorable and remaining constant, cyclone genesis should be directly related to the moist buoyancy potential or the magnitude of the boundary layer to middle troposphere difference of Θ_e. This buoyancy is well specified by the Θ_e difference between the surface and the 500 mb level.

4.6 Role of Middle Troposphere Humidity

It is observed that tropical cyclones form only in regions where the seasonally averaged values of middle level humidity are high. Cyclones do not form where seasonal values of middle level humidity are low.

Other factors being equal, an environment of high middle level humidity is more conducive to deep cumulus convection and greater vertical coupling of the troposphere than is a dry middle level environment. High middle level humidity is also conducive to high cloud precipitation efficiency (Lopez, 1973).

It must be remembered that in cyclone genesis situations the oceanic boundary layer does not experience a diurnal warming cycle as is present over land and that tropospheric vertical wind shears are small. These conditions preclude squall line development and intense buoyant up-and-downdraft activity that is possible where large diurnal heating cycles and strong tropospheric wind shears are present. Although deep and intense Cb convection occurs over middle latitude land areas in conditions of low middle level relative humidity, Cb convection does not typically occur over ocean regions when middle level humidity is less than 50-60 percent (Ruprecht and Gray, 1974). Over tropical oceans high middle level vapor content appears to be a strong enhancer rather than an inhibitor of deep cumulus convection.

To better understand the influence of middle level humidity on Cb parcel buoyancy, Table 3 has been prepared. This table compares the temperature of a parcel lifted moist adiabatically from 850 to 250 mb under assumed entrainment rates of 0, 10, and 20 percent per 100 mb in two environments whose humidity above 850 mb is different by 25 percent. An initial parcel buoyancy of 1.7^{o} at 850 mb is assumed for both environments. Parcels entrain air with the humidity values of the environment listed at the left of this table. The large influence of middle level humidity differences of only 25 percent in reducing updraft parcel buoyancy is evident when entrainment is occurring.

High middle level humidity also leads to increased precipitation efficiency and the greater likelihood of enthalpy increase. Lopez's (1973) model of the cumulus life cycle indicates a direct relationship between cumulus induced precipitation at the ground and upper level relative humidity. When upper level humidity is low, much of the cumulus induced condensation does not fall to the ground but is reevaporated. This produces upper level cooling which can act to suppress rather than enhance the increase of enthalpy.

It will thus be assumed that cyclone genesis is directly related to the average of 500 and 700 mb relative humidity. The seasonal values of middle level humidity are related to a humidity parameter which is assumed to be associated with cyclone genesis. This humidity parameter varies from 0 to 0.5. Cyclone development is not possible if seasonal 500-700 mb humidity is less than 40 percent. This factor increases linearly from 0 for seasonal humidity $\leq$ 40% to 0.5 for seasonal humidity

Table 3

Influence of Middle Troposphere Humidity on Parcel Buoyancy for Different Values of Entrainment from 850 mb.

Cumulus Parcel Temperature Minus Environment Temperature (^{o}C)

Pressure Level (mb)	Relative Humidity (percent)		No Entrainment		For Environment Entrainment of 10% per 100 mb		For Environment Entrainment of 20% per 100 mb	
	Typical Cluster Humidity	Typical Cluster Humidity less 25%	Higher Humidity	Lower Humidity	Higher Humidity	Lower Humidity	Higher Humidity	Lower Humidity
850	80	55	1.7	1.7	1.7	1.7	1.7	1.7
750	78	53	2.2	2.2	1.5	0.8	0.6	-1.1
650	76	51	3.0	3.0	1.7	0.4	0.2	-2.7
550	74	49	3.8	3.8	2.0	0.2	0.1	-3.7
450	72	47	4.2	4.2	2.0	-0.1	0	-4.6
350	70	45	4.0	4.0	1.6	-0.5	-0.7	-5.6
250	65	40	4.0	4.0	1.5	-0.7	-0.9	-6.0

$\geq 70\%$. Thus, the RH factor is specified as

0.5 for RH $\geq 70\%$
0.4 for RH 55-69%
0.2 for RH 46-54%
0.05 for RH 40-45%
0.0 for RH <40%

Figure 20 portrays seasonal values of 500-700 mb relative humidity.

Conclusion. Other conditions being favorable and remaining constant, tropical cyclone genesis frequency should be directly related to seasonal values of middle tropospheric humidity. The relationship between middle level moisture and cyclone genesis is expressed by the above relative humidity factor.

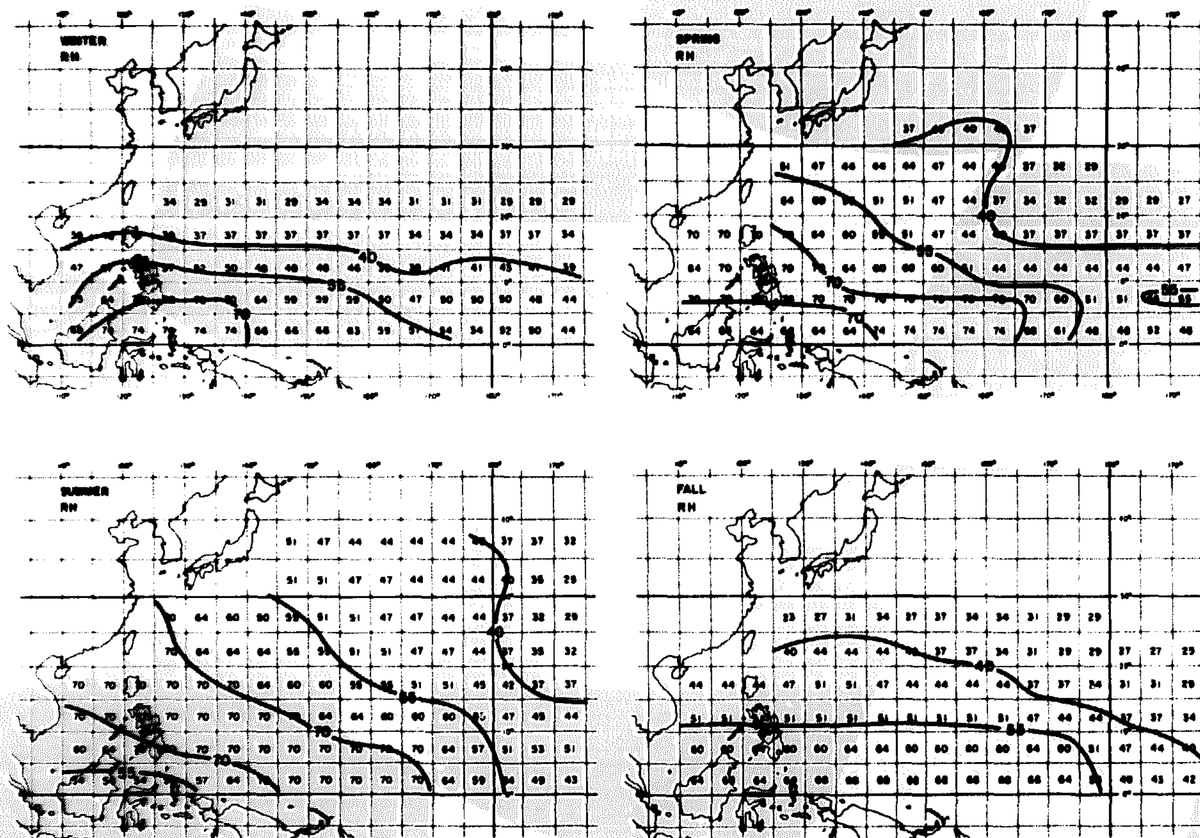


Fig. 20. Seasonal average of 5° square 500 to 700 mb average relative humidity.

5. SPECIFICATION OF SEASONAL GENESIS FREQUENCY

It is hypothesized that tropical cyclone formation will be most frequent in the regions and in the seasons when the product of the six primary genesis parameters just discussed are a maximum. Seasonal Genesis Frequency should be determined by the product of:

$$(\zeta_r)\ (f)\ (1/S_z)\quad (STF)\left(\frac{\partial \Theta_e}{\partial p}\right)\ (\overline{RH})$$

where ζ_r = Relative vorticity at top of planetary boundary layer or ~950 mb

f = Coriolis parameter

$1/S_z$ = $1/\left|\partial \mathbf{V}_h/\partial p\right|$ or the inverse of the absolute value of the vertical shear of the horizontal wind between 950 and 200 mb

$$STF = \frac{T_{SST} - 79^{o}}{1 + (T_{SST} - T_{200\ ft})}$$

$\frac{\partial \Theta_e}{\partial p}$ = Vertical gradient of Θ_e between the surface and 500 mb

$\overline{RH}$ = Mean relative humidity between the levels of 500-700 mb

Zero and negative values of the above parameters indicate no genesis potential.

There is some difficulty with using these seasonal values directly. Seasonal values are not a close measure of what the daily parameter values can be. Thus, the seasonal relative vorticity in the western Atlantic is slightly negative. The above seasonally determined genesis potential would not predict cyclone genesis in a region where it obviously occurs. When the seasonal values of $\left|\partial \mathbf{V}_h/\partial p\right|$ are small, unreasonably large values of $1/S_z$ are obtained, which are not representative of the average of the daily values. Similarly the seasonal values of

$\partial\Theta_e/\partial p$ between the surface and 500 mb could be zero or negative, but individual time period values can be positive.

To cover the range of possible daily deviations of three of these parameters, arbitrary units were added to the seasonal values to simulate in an approximate sense what the average of the positive daily deviations of these values could be. Thus, five units of $10^{-6}sec^{-1}$ vorticity were added to all the seasonally measured values of this parameter and $5^{o}K$ was added to all the seasonal values of $\partial\Theta_e/\partial p$. To prevent unreasonably large values of the genesis frequency when $\left|\partial V_h/\partial p\right|$ approaches zero, and also to assure that the seasonal average of $\left|\partial V_h/\partial p\right|$ is more representative of the seasonal average of the daily values, 5 m/sec was arbitrarily added to all the seasonal vertical shear values. The minimum value of the vertical shear parameter is thus 5 units.

A Seasonal Genesis Parameter is now defined as

$$\begin{pmatrix}\text{Seasonal}\\\text{Genesis}\\\text{Parameter}\end{pmatrix} \propto \left[\begin{pmatrix}\text{Vorticity}\\\text{Parameter}\end{pmatrix}\begin{pmatrix}\text{Coriolis}\\\text{Parameter}\end{pmatrix}\begin{pmatrix}\text{Vertical Shear}\\\text{Parameter}\end{pmatrix} \text{ X } \begin{pmatrix}\text{Sea Temperature}\\\text{Factor}\end{pmatrix}\begin{pmatrix}\text{Moist Stability}\\\text{Parameter}\end{pmatrix}\begin{pmatrix}\text{Humidity}\\\text{Parameter}\end{pmatrix}\right] \quad (9)$$

where

$\begin{pmatrix}\text{Vorticity}\\\text{Parameter}\end{pmatrix}$ = $(\zeta_r + 5)$, where ζ_r is determined in units of $10^{-6}sec^{-1}$.

$\begin{pmatrix}\text{Coriolis}\\\text{Parameter}\end{pmatrix}$ = f or $2\Omega\sin\varphi$, where Ω is the rotation rate of the earth and φ denotes latitude.

$\begin{pmatrix}\text{Vertical Shear}\\\text{Parameter}\end{pmatrix}$ = $1/(S_z+5)$ where $S_z = \left|\partial V_h/\partial p\right|$ is determined in units of m/sec per 750 mb.

$\begin{pmatrix}\text{Sea Temperature}\\\text{Factor}\end{pmatrix}$ = STF or $\dfrac{T_{SST} - 79^{o}F}{1 + (T_{SST} - T_{200})}$

(Moist Stability Parameter) = $\partial\theta_e/\partial p + 5$, where $\partial\theta_e/\partial p$ is in units of $^{\circ}$K per 500 mb.

(Humidity Parameter) = $\overline{RH}$ which varies from 0-0.5. $\overline{RH}$ = 0.5 for mean 700-500 mb humidity $\geq$ 70%, 0.4 for mean humidity of 55-69%, 0.2 for humidity of 46-54%, .05 for humidity of 40-45% and 0 for values < 40%.

Table 4 summarizes these parameters and their role in cyclone genesis. This Seasonal Genesis Parameter may also be thought of in the form of

$$\begin{pmatrix}\text{Seasonal}\\ \text{Genesis}\\ \text{Parameter}\end{pmatrix} = (\text{Dynamic Potential}) \times (\text{Thermal Potential}) \qquad (10)$$

where

$$\text{Dynamic Potential} = (f)\ (\zeta_r + 5)\ 1/(S_z + 5) \qquad (11)$$

$$\text{Thermal Potential} = (\text{STF})\ (\partial\theta_e/\partial p + 5)\ (\overline{RH}\ \text{Parameter}) \qquad (12)$$

Dynamic Potential is expressed in units of 10^{-12}sec^{-2} per m/sec per 750 mb and Thermal Potential in units of $^{\circ}$K per 500 mb. Thermal Potential may also be thought of as the potential for Cb convection.

Dynamic Potential. Seasonal values of the dynamic potential or $(\zeta_r + 5)$ (f) $(1/(S_z + 5))$ in units of 10^{-12}sec^{-2} per m/sec per 750 mb are portrayed in Fig. 21. The correlation between high values of this potential and the location and frequency of seasonal cyclone genesis has a number of significant shortcomings. Cyclone frequency is related to more than dynamic influences. Thermal cumulus buoyancy considerations must also play an important role.

Thermal Potential. Figure 22 gives seasonal values of the product of the Sea Temperature Factor (STF) - Fig. 18, the moist stability $(\partial\theta_e/\partial p + 5)$

Table 4

Summary of Primary Genesis Parameters

Parameter	Favorable Condition	Genesis Role
1. Vorticity Parameter -- $(\zeta_r + 5)$ at $\sim$950 mb where ζ_r is in units of $10^{-6}sec^{-1}$	Large	Produce necessary low level mass, moisture, and momentum convergence through Ekman-type boundary layer friction
2. Coriolis Parameter -- or f	Large	Allow for pressure gradient and sustaining of boundary layer winds against frictional dissipation
3. Vertical Shear Parameter -- $1/(S_z + 5)$ where $1/S_z$ is in units of $(m/sec)^{-1}/750$ mb	Large	Allow condensation warming to be concentrated over moving disturbance; i.e. inhibit 200-500 mb ventilation energy
4. Sea Temperature Factor (STF) -- or $\frac{T_{SST} - 79^{o}F}{1 + (S_{SST} - T_{200})}$	Large	Maintain surface Θ_e values in conditions of strong winds, upwelling, and large sea to air energy transfers
5. Moist Stability Parameter -- $(\frac{\partial\Theta_e}{\partial p})$ from surface to 500 mb	Large	Permit Cb cumulus convection
6. Humidity Parameter -- $\overline{RH}$ between 500 and 700 mb	Large	Allow for deep cumulus convection and high rainfall efficiency

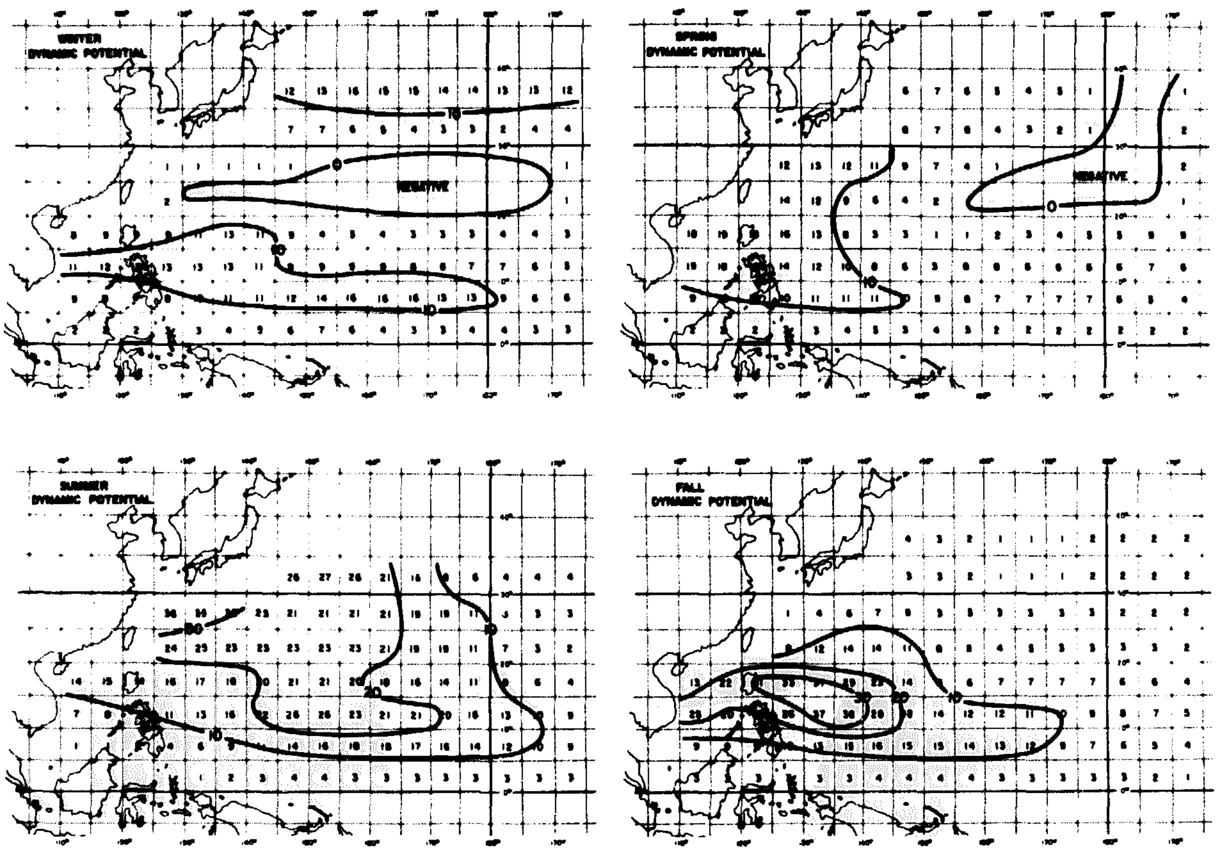


Fig. 21. **Seasonal averages of 5° square Dynamic Potential or ($\zeta_r + 5$) (f) ($1/[S_z + 5]$) in units of (10^{-12} sec^{-2}/m/sec per 750 mb)**

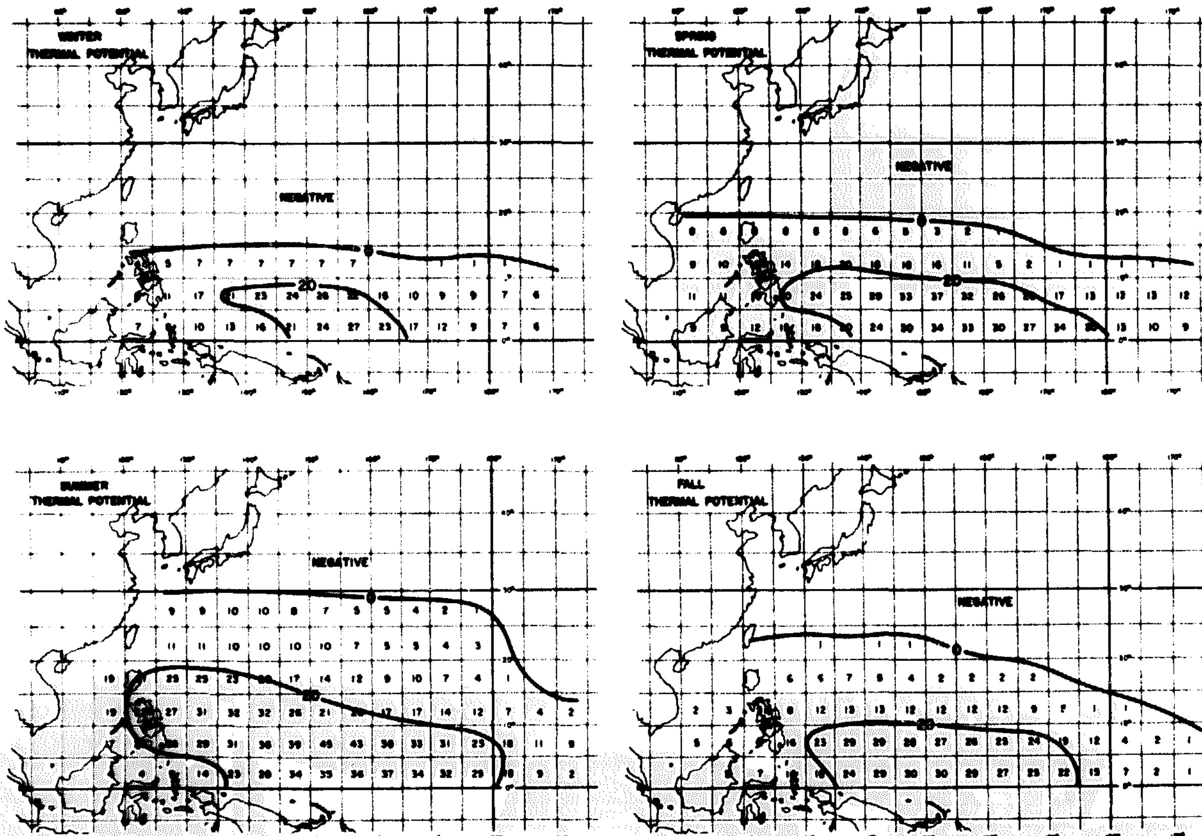


Fig. 22. Seasonal values of 5° square Thermal Potential or $(\Delta\Theta_e + 5)$x(STF)x(RH factor) in units of $^{\circ}$K per 500 mb.

- Fig. 19, and relative humidity factor derived from the humidity values in Fig. 20. Isolines are portrayed in values of ^{o}K per 500 mb. This product may be considered as the thermal potential, or the potential for Cb convection. Note the sharp northward cut-off in the sea temperature factor. This potential restricts genesis to latitudes $<15^{o}$ in the winter and inhibits genesis poleward of 30^{o}N in general. Equatorwards of 15^{o} the sea temperature factor is favorable in all seasons.

The moist stability parameter $(\partial\Theta_e/\partial p + 5)$ is generally large in all oceanic locations except in the sub-tropical latitudes in winter. Cyclone genesis will not generally occur when the magnitude of this parameter is less than 15. When above 15, variations of this parameter do not greatly affect spatial and seasonal variations of genesis. Seasonal variation of the 15 line rather well specifies the poleward extent of cyclone genesis.

Seasonal values of the humidity parameter are also very important in specifying the spatial and seasonal frequency of cyclone genesis. Cyclones do not form in regions where the seasonal 500-700 mb relative humidity is less than 40 percent. This parameter is a major factor in specifying differences in winter and summer cyclone frequency at low latitudes.

Verification. Figure 23 portrays seasonal values of the (Dynamic Potential) x (Thermal Potential) or the forecast potential. This has units of 1.5×10^{-10} sec^{-2} ^{o}K per m/sec. When expressed in these units this gives just the right factor required to specify seasonal forecast cyclone genesis frequency in number per 5^{o} latitude-longitude square per 10 years. A small degree of smoothing has been performed. Figures 24 and 25 compare the seasonal forecast potential and observed cyclone

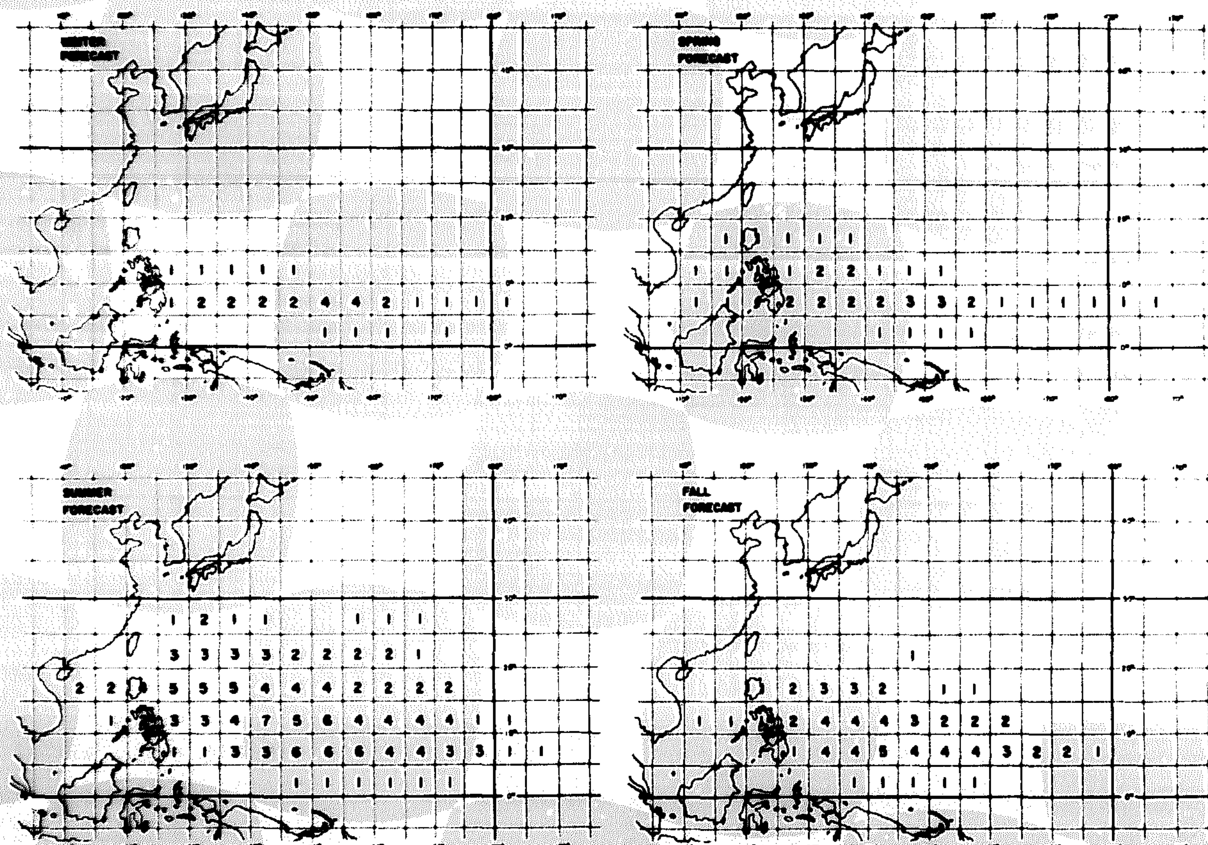


Fig. 23. Seasonal average of 5° square <u>forecast potential</u> or $\left(\begin{smallmatrix}\text{Dynamic}\\ \text{Potential}\end{smallmatrix}\right) \times \left(\begin{smallmatrix}\text{Thermal}\\ \text{Potential}\end{smallmatrix}\right)$ in units of 1.5×10^{-10} sec^{-2} ^{o}K per m/sec. Values shown are of cyclone forecast number per 5° Marsden square per 10 years.

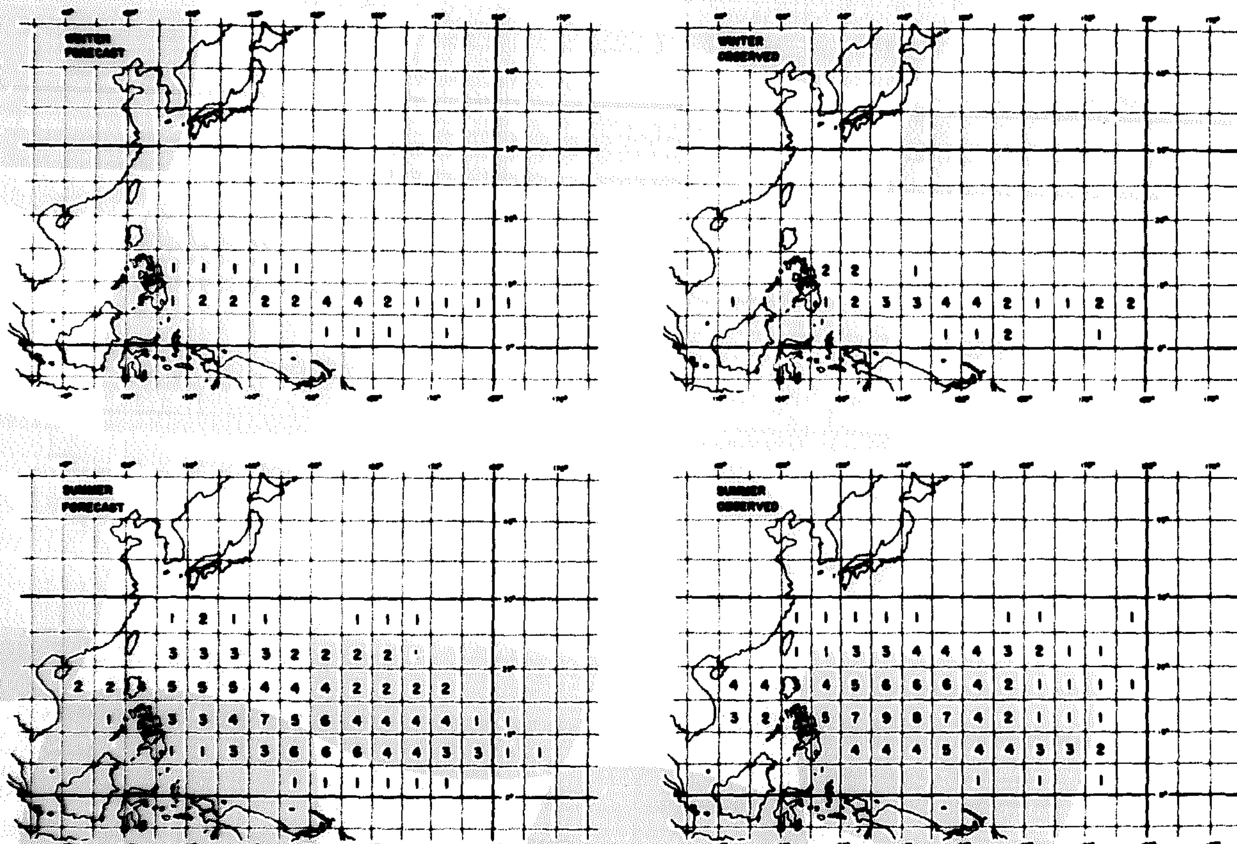


Fig. 24. Comparison of forecast (left) and observed (right) number of tropical cyclone initial genesis points per 5° Marsden square per 10 years during the winter and summer seasons.

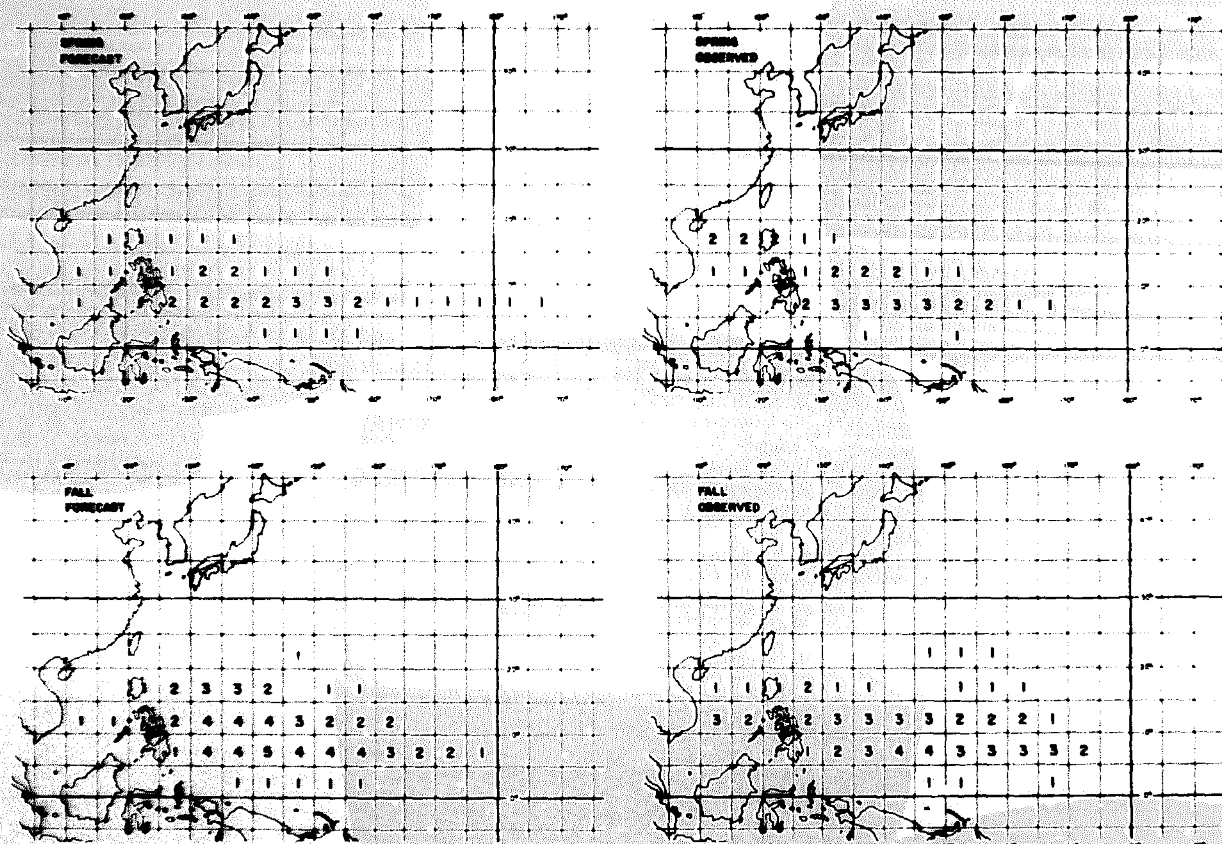


Fig. 25. Comparison of forecast (left) and observed (right) number of tropical cyclone initial genesis points per 5° Marsden square per 10 years during the spring and fall seasons.

genesis frequency in number per 5^{o} square per 10 years. Figure 26 stratifies seasonal western Pacific forecast potential and observed cyclone genesis by 5^{o} latitude segments. The very close correspondence between predicted and observed seasonal cyclone frequency lends support to the earlier arguments concerning the relevant physical genesis parameters. It was not expected that the agreement between predicted and observed genesis frequencies would be this close.

Application to Daily Forecasting. It is hoped that these seasonal genesis statistics can be adapted for use on a daily basis at the various government offices responsible for tropical cyclone monitoring. As the satellite and jet aircraft wind measurements become more frequent and accurate, better daily measurements of low level vorticity and tropospheric vertical shear will be obtained. This should lead to improvement in the daily measurement of the genesis potential and also, hopefully, in the improved forecasting of individual cyclone genesis.

LATITUDE	WINTER		SPRING		SUMMER		FALL		ANNUAL	
	OBSERVED	FORECAST	OBSERVED	FORECAST	OBSERVED	FORECAST	OBSERVED	FORECAST	OBSERVED	FORECAST
25°–30°					8	8			8	8
20°–25°					27	21	3	1	30	22
15°–20°			8	5	48	43	10	13	66	61
10°–15°	5	5	11	11	53	49	28	26	97	91
5°–10°	27	24	20	25	33	42	28	34	108	125
0°–5°	5	4	2	4	3	6	3	5	13	19
TOTAL	37	33	41	45	172	169	72	79	322	326

Fig. 26. Comparison of seasonal and annual 10-year total of observed and forecast initial cyclone genesis points by latitude.

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APPENDIX

Data Sources for Various Calculations.

1. Seasonal vorticity was calculated from the monthly gradient level streamline charts of G. D. Atkinson and J. C. Sadler's 1970 report entitled "Mean Cloudiness and Gradient Wind Charts over the Tropics", (Air Weather Service Technical Report No. 215, Vol. II). Vorticity was calculated for each month and then averaged to obtain seasonal mean values.

2. Vertical shear of the horizontal wind from 950 to 200 mb was calculated from the low level data source listed above and from specially constructed 200 mb flow fields for the months of February, May, August, and November which were accomplished by the author at Colorado State University. The specially constructed 200 mb maps were made from the NOAA Monthly Climatic Data for the World where 200 mb monthly vector winds are listed. It was assumed that the February, May, August and November 200 mb wind vectors were typical of those of the January-March, April-June, July-September, and October-December periods respectively. The vertical shear charts represent the magnitude of the mean wind vector differences between the 200 and 950 mb levels.

3. The Sea Temperature Factor calculations were made from the ocean temperature depth data of the North Pacific as reported by Robinson and Bauer (1971).

4. Θ_e gradients between the surface and 500 mb were calculated from monthly values of temperature and moisture from the NOAA Monthly Climatic Data of the World listings for the months of February, May, August, and November. These months were taken as representative of

the January-March, April-June, July-September, and October-December seasonal averages.

5. Relative humidity at 500 and 700 mb was obtained from the upper air monthly mean moisture information contained in the NOAA Monthly Climatic Data of the World. This data was adjusted and smoothed to fit as well as possible the upper air moisture data contained in the NAVAIR report of Crutcher and Meserve (1970). Some smoothing and reconciliation of these data sources was required.

6. Initial tropical cyclone genesis location data was obtained from the author's previous global data on tropical cyclone origin and from undated information listed in the combined U.S. Navy Environmental Prediction Research Facility and the NOAA National Weather Records Center, Asheville, NC, printout tape of tropical cyclone data, position, intensity, etc. This tape was kindly furnished to the author by Mr. Samson Brand of the U.S. Navy's Environmental Prediction Research Facility.

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